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STUDIES OF HEAVY DETRITAL MINERALS

By C. Osborne Hutton

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Abstract

This paper embodies the results of a detailed investigation of the heavy detrital minerals in the sands and gravels in streams, beaches, terraces etc. in a wide variety of localities in South Island, New Zealand.

Appropriate techniques using screens, bolting silk, electromagnets, and the centrifuge among other devices have been developed for this and earlier work, and some well-known techniques have been modified to suit the special conditions imposed during this investigation. The centrifuge proves superior to the ordinary separating funnel if fine-grained material is handled. Various acids prove effective as solvents for calcium carbonate. Attention has been focussed on the inadvisability of using Clerici solution under certain conditions. Heavy mineral fractionation by heavy liquids in flat-bottomed porcelain dishes, rather than in ordinary filter funnel or separating funnel, is fully described, as well as the preparation of pure mineral samples.

Detailed optical, density, and chemical studies of over 50 placer minerals are fully reported. Geological occurrence of the economically important tantalian cassiterite, gadolinite, monazite, thorite and its varieties, xenotime, scheelite, and zircon and its varieties is emphasized. Some varieties of thorite exhibit properties that are absolutely distinct from those previously recorded.

The nature and causes of the metamict state in radioactive minerals are considered in the light of modern data on crystal structure, and although this question is not fully answered the opinion is ventured that bombardment of unstable structures by alpha-particles is a vital factor. Micro-fissuring of radioactive minerals is discussed and it is concluded that the web of fracture therein may be attributed to expansion resulting from transformation to the metamict state.

An inquiry into the location and nature of the source rocks of the heavy detrital minerals concludes the report.

Acknowledgments

The writer wishes to express his appreciation to Mr. F. T. Seelye for excellent analyses carried out with so much care and for numerous discussions on chemical problems; to Professors Adolf Pabst and F. J. Turner of the University of California and Professor D. L. Webster, Stanford University for valuable arguments and constructive criticism; to a number of gold dredging companies, Mr. Ian MacKellar, and many others for their kindness and assistance in supplying material; to Professor F. W. Tickell, Stanford University, for his painstaking editing; to Miss Louise Ricks for her accurate typing of the manuscript; and to Mr. Frank D. Banham for his help, co-operation, and advice in the fabrication of some of the special instruments used in this study.

Any acknowledgment would be incomplete without an expression of deep gratitude to my wife, May, for her encouragement and active help at all times.

The author alone is responsible for errors of fact or interpretation.

Object of the Work

The investigation was begun as part of a program involving the search for radioactive minerals in territory from which they had not hitherto been reported. As the work progressed, the writer set about attempting to determine

the nature of the heavy detrital minerals in a wider range of localities with particular reference to their physical and chemical properties. To gain some idea of the ultimate source rocks of some of the more important minerals, the minor accessory constituents of a wide variety of rocks from appropriate localities were separated and studied.

Scope of the Work

In the main, the work involved a detailed study of the heavy detrital minerals in beach, river, terrace, and dune deposits in a wide range of localities in South Island, New Zealand, with particular attention to material from distributive provinces in which acid plutonic rocks outcropped. Concentrates from dredges, prospectors' and gold-miners' pannings and washings, pay-streaks on beaches and along water-courses, and the products of the writer's pannings in a wide range of localities where dredges or prospectors were not active formed the basis of the investigation.

Appropriate methods were devised to examine the material, which in most instances contained a preponderance of heavy minerals, either ferromagnetic or strongly to moderately paramagnetic. At the same time, the efficacy of various laboratory methods of mineral concentration and preparation was investigated, particularly techniques for preparation of pure mineral fractions required for analytical purposes. In view of the difficulties involved and the scanty work on such problems, the diagnosis of opaque detrital minerals was considered.

Several unusual minerals were diagnosed by optical means alone, but in view of the poor optical and chemical data available for some uranium- and thorium-bearing minerals and the various rare earth compounds, all information possible on the chemical compositions of pure minerals, especially the thorites, was sought to enable an understanding of the optical data.

Although the general problem of the nature and distribution of the heavy detrital minerals throughout South Island is incomplete, the writer's departure from New Zealand will not permit his undertaking any further studies on this problem; however, the work should indicate the directions for future investigations and it is hoped that it may serve as a guide in general geological problems, particularly on the west coast of South Island. Further, it may show geologists the value from many points of view, economically, mineralogically etc., of work of this nature.

Previous Work

In New Zealand, very few special studies of the heavy mineral assemblages of beach and stream sands, gold-miners' and dredge concentrates, or incoherent rocks have been made, and only a few of these are in any detail (Finch, 1948; Hutton, 1940b; 1945a; 1945b; 1947a; 1947b; Hutton and Turner, 1936; Morgan and Gibson, 1927; Speight, 1928; N. H. Taylor, 1933; Turner, 1944; Williams, 1934; and Wylie, 1937); most of these investigations have been concerned with titaniferous ironsands on the west coast of the North Island. No thorough studies of South Island Quaternary sediments have been made although several New Zealand Geological Survey Bulletins discuss the South Island west coast blacksand beaches, chiefly on account of their gold content. In some of these publications and in numerous Annual Reports of the Dominion Laboratory, references are made, sometimes without sufficient data, to the presence of monazite, cassiterite, platinum, osmiridium, yttrotantalite etc. in auriferous placers and other deposits (Bell and Fraser, 1906; Henderson, 1917; Morgan, 1908; 1911; 1927; and Webb, 1910). Analyses of concentrates from sluicing claims, dredging companies, and other sources (material from Addison's, Westport, Dominion Analyst, 1917, p. 25), and blacksand paystreaks (for material from Charleston see Dominion Analyst, 1926, p. 21) have been recorded but these data are usually unsupported by mineralogical determinations.

Part 1. Mineralogical Techniques

Introduction

Some of the methods employed by the writer in this and other work are recorded in detail for the use of others concerned with similar problems, but some of the methods discussed are not original, having been adapted or modified for some specific purpose. Since the various prob-

FIGURE. 1—LOCALITY MAP OF SOUTH ISLAND, NEW ZEALAND

lems concerned with mineral isolation, concentration, determination, and so on differ so much, it is neither easy nor desirable to lay down definite rules for work of this nature, and therefore the experimenter must devise the most appropriate combination of manipulations for a specific task.

Preparation of Original Material for Heavy Mineral Separation

The original materials can be grouped into four main categories:

(a). Soft, coherent but not dominantly carbonate-bearing rocks.

(b). Hard rocks devoid of much carbonate.
(c). Carbonate-rich rocks.
(d). Incoherent materials.

Specimens belonging to categories (a)–(c) require complete disaggregation; for this purpose specimens weighing about 2–3 pounds have usually been found to be sufficient.

(a). A soft specimen may be completely broken down by working small fragments between the fingers; if not amenable to this treatment, it may be so after soaking or boiling in water. If a rock will not yield to these methods it should be soaked in a hot solution of sodium sulphate or ordinary photographers' hypo, until the specimen itself is thoroughly saturated. It should then be removed from the solution, placed in a suitable dish, and the salt allowed to crystallize. Expansion of the growing crystals will often reduce a coherent rock to fine composite particles more easily dealt with. Some rhyolite tuffs were thus broken down to almost completely monomineralic powders. If these methods fail, the rock should be broken into small fragments and finally rubbed down, but not ground, in a cast iron mortar with a long-handled iron-headed pestle.

(b). Hard rocks may be broken down to conveniently sized fragments in a jaw crusher or placed in a deep iron mortar and pounded with the long-handled pestle into small chips. For disaggregation of the fragments pounding them in an iron mortar may be satisfactory in the case of specimens that are semi-coherent such as sandstones and some schists, but with fresh, massive igneous rocks, quartzites, etc., disintegration may be conveniently accomplished with a special percussion steel mortar and pestle (Fig. 2). The eccentrically placed percussion point is a necessary feature of this tool since it serves to prevent wedging of the pestle in the sleeve by the powder forming from the rock chips under treatment.

It is necessary to screen the crushed material of the last two groups, but the choice of sieve openings depends very much on the grain sizes of the rocks concerned. The writer has found it convenient to screen a rock powder through sieves with openings of 4, 2, 1, 0.5, and 0.25 mm. (Wentworth grade scale) and to examine test samples from each fraction beneath a binocular microscope for composite particles. If these are found, almost certainly in the coarser screenings, the fraction or fractions containing them are reduced further. When sizing of a disaggregated sediment is required,

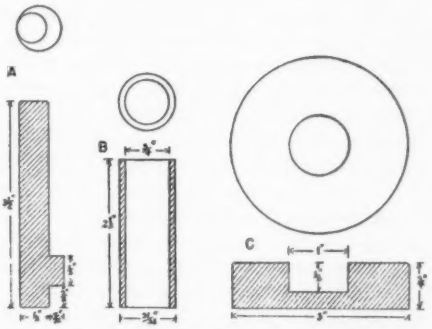

FIGURE 2.—PLAN AND SECTION OF PERCUSSION STEEL PESTLE AND MORTAR
A. pestle; B. sleeve; C. mortar.

screens with openings 0.125 and 0.062 mm. should be added. However, if it is only proposed to investigate the minor accessories of a rock it is preferable to use bolting silk for the finer particle sizes since this is readily cleaned and contamination is less likely than with fine mesh metal screens. A nest of silk screens is used, details of which will be given later.

After disaggregation is complete, the powder must be freed from clays and dust before satisfactory sedimentation or electromagnetic fractionation can be undertaken. This may be accomplished by careful elutriation of 50 gm. portions of rock powder in a pear-shaped separatory funnel under a slow but constant up-flow of water with simultaneous collection of the overflow in a series of 1- or 2-liter capacity beakers, if ultimate study of the clay or fine fraction is desired. It is just as efficient and certainly much quicker, to pour the powder into a 1-liter beaker which is nearly filled with water making certain, at the same time, that the powder is thoroughly wetted and dispersed; with very fine powders, dampen the powder with alcohol before adding water to hasten thorough wetting. After the muddy liquid settles for about 90 seconds, 750 ml. of the

supernatant liquid are either decanted or siphoned off; this liquid is retained as before. The process is repeated until, after a 90-second settling period, the supernatant liquid no longer remains cloudy.

(c). Carbonate-rich rocks should be broken down by any suitable method, the powder so obtained washed free from clay and fines, followed by solution in acid to remove carbonates. Since it is generally held that a number of heavy minerals found in carbonate and other rocks are soluble in strong mineral acids, a number of authors have suggested the use of either citric or acetic acids instead. The present writer believes that the usual constituents of a heavy mineral assemblage are almost insoluble when treated with mineral acids of moderate concentration in the cold for short periods of approximately 20–30 minutes, and this also applies to apatite except in a very finely divided form, less than about 50 microns. A similar view expressed by R. D. Reed (1924, p. 324) appears to have been lost sight of. A greater disadvantage of the use of strong mineral acids such as HCl on detrital mineral grains than the negligible possibility of loss of essential material, is the not infrequent production of etch figures, and if these are found on grains treated in this manner they cannot readily be distinguished for natural ones. However, neither acetic nor citric acid is at all satisfactory when large quantities of carbonates are to be dissolved, owing to the limited solubilities of their salts[1], particularly calcium citrate; in addition, both the calcium acetate and citrate become less soluble with temperature increase in marked contrast to the solubility of calcium chloride. Nevertheless acetic acid is useful when calcium carbonate is only a minor constituent of a sediment[2]. Instead about 4N HCl has been found most satisfactory in spite of possible partial solution of some of the heavy mineral suite. The procedure usually is as follows: A 25 gm. portion of powdered carbonate-rock is placed in a 400 ml. capacity beaker and treated with about 150 ml. of 4N HCl, added 50 ml. at a time; normally this should be sufficient acid to dissolve all of the carbonate present. When effervescence has ceased the liquid is heated to ensure complete solution of any dolomite present. After settling of the insoluble residue, the supernatant liquid is carefully decanted through a filter, and 20 ml. of 4N HCl added to the residue to prove that solution of the carbonates has been complete. The residue is then thoroughly washed with distilled water and dried in an oven at 90°C.

Pure HCl has been suggested for solution of carbonates since commercial type acid contains SO_4 ions and this would result in the contamination of the insoluble residue with calcium sulphate, presumably in the form of gypsum. Because analytically pure HCl is much too expensive for this use, commercial acid is recommended; but since a small check portion of the original powdered carbonate rock without prior acid treatment should be fractionated in bromoform of density about 2.88 to detect the presence of any acid soluble minerals other than carbonates therein, the presence of introduced gypsum can be observed. Even if the specimen should be contaminated with gypsum, and no instance of this has occurred in the writer's work, the heavy mineral fraction should be devoid of this constituent if the separation has been clean.

(d). Incoherent materials studied by the writer include a wide variety of eluvial, alluvial, littoral, and aeolian materials, and in this investigation a wide range of tailing, sludges, and panned concentrates was also examined.

Careful elutriation is often the sole requirement for preparing specimens of incoherent materials for mineral fractionation; with dredge concentrates and pannings, not even this treatment is required. However, tailings, panned concentrates as well as some sands contain a high proportion of minerals of density greater than 2.85, and it is more satisfactory to fractionate electro-magnetically each grade-size before flotation in heavy liquids.

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[1] Solubility of calcium acetate: 2.72 gms. in 100 gms. of water at 0°C. and 2.29 gms. in 100 gms. of water at 100°C.

Solubility of calcium citrate: 0.22 gms. of $Ca(C_6H_5O_7)$ per 100 cc. of aqueous solution saturated with $Ca(C_6H_5O_7)\cdot 4H_2O$ at 30°C. and at 95°C. 0.183 gms. of $Ca(C_6H_5O_7)$ per 100 cc. of aqueous solution (Seidell, 1940, pp. 246, 255).

[2] Professor F. G. Tickell has suggested to the writer that trichlor-acetic acid, a strong organic acid, might prove useful for dissolving carbonates.

Fractionation in Heavy Liquids

Both bromoform and acetylene tetrabromide have been used but the writer favors bromoform because it has a decidedly lower viscosity, *viz.* 0.068 as compared with 0.098 poise for $C_2H_2Br_4$, and hence filtration and sedimentation are both quicker, a factor of some importance when a large number of specimens has to be run through. In the writer's opinion this advantage outweighs that of slightly higher density of acetylene tetrabromide for most purposes. Alcohol, acetone, or benzene (C_6H_6) are satisfactory diluents or wash liquids and bromoform from washings is easily recovered. Both alcohol and acetone are readily removed from a mixture of either of these compounds with bromoform by washing with liberal quantities of water and drying the recovered bromoform over fused calcium chloride, whereas benzene must be removed either by fractional distillation or by fractional crystallization. If fractional distillation is employed, the entire operation should be carried out at reduced pressure (approximately 16–25 mm. of Hg) and preferably on a shaded bench, but certainly out of direct sunlight if extensive decomposition of bromoform is to be prevented. If methylene iodide is also present in this mixture it may be recovered at the same time although the percentage recovery is not as good as that for bromoform. A more convenient method of recovery, by which little decomposition need occur, is to place the benzene-bromoform mixture in a flask in a freezing mixture; since the melting point of bromoform is about 7°C. and benzene 5.5°C., bromoform will crystallize out before the benzene commences to do so. The same procedure may be adopted with bromoform-alcohol mixtures.

The method used for the actual separation of the heavy mineral fraction will depend upon the average grain-size of the original material. For grains finer than 0.125 mm. in diameter the usual separating funnel methods are unsatisfactory, and such material must be separated centrifugally (Hutton, 1943). This opinion is not in agreement with Grout's (1937) findings on the efficacy of centrifuge separations, but it is in partial support of the data presented by Tyler and Marsden (1937). Additional work on this problem, since the publication of the author's earlier study, has shown beyond all doubt that for very fine-grained sediments both the yield of heavy minerals and the relative frequencies of the different minerals are significantly different when obtained by centrifuge methods instead of the usual funnel technique.

If the material is relatively coarse, as are most panned and dredge concentrates, fractionation is readily and accurately accomplished in a simple porcelain evaporation dish about 3–3½ inches in diameter, preferably one with vertical sides, such as a flat-bottomed Coors dish.

The dish is half filled with bromoform, or what liquid the operator wishes to use, and about 20–25 gm. of cleaned, graded, and absolutely dry material is carefully poured into the liquid. The grains floating at the surface are continuously but carefully stirred until fractionation is considered complete. Then the dish is tilted slightly and the layer of heavy grains is brushed towards the side of the dish below the lip; this operation may be assisted by tapping the bottom of the dish so that the grains roll or slide. Now wash back into the dish with bromoform from a wash-bottle the wet but mainly light grains that are adhering to the brush and carefully stir the surface layer once again, with special care not to cause convection currents strong enough to bring about redistribution of the segregated heavy grains. After 3–4 minutes, carefully skim aside a portion of the light layer and determine whether there has been further sedimentation. If not, the light fraction is recovered by decantation through a filter[3] and the dish and heavy fraction contained therein carefully washed with bromoform, decanting the liquid after each washing until only the heavy fraction, completely free from light grains, remains in the dish. Wash finally with alcohol, decant wash-liquid, and place the dish with the "heavies" on a hot plate at a temperature of 90°C. If it is necessary to separate additional portions of the original material a supplementary 20–25 gm. lot should be added after decantation of the first light

[3] Whatman's No. 4 papers have been found most suitable for this work.

fraction; washing with alcohol is carried out after the last separation has been completed.

Using an open dish is a relatively rapid way to fractionate large amounts of material and the wide surface area and absence of sloping sides or shoulders assists rapid sedimentation. Washing the heavy residue with bromoform in the dish after decantation of the bulk of the light grains provides one with a very clean fraction, uncontaminated by light grains, and final recovery of the "heavies" in the dish rather than in a filter-paper means a minimum of loss of very small and fine-grained residues. Finally, porcelain dishes are cheaper to buy and easier to clean than glass separatory funnels.

For separating heavy minerals from fine-grained materials by centrifuging methods, modified Taylor (G. L. Taylor, 1933) tubes are used. Two or four tubes are used simultaneously and each tube has a total capacity of about 110 ml. with a narrow waist dividing each tube into a small lower bulb, with a capacity of approximately 6–10 ml., and a larger upper portion. This enables a large residue to be accumulated before it is necessary to wash out the heavy fraction. Samples are usually centrifuged for from 2–5 minutes, at speeds ranging from 1500–3500 revolutions per minute.[4] The method is as follows: To each thimble or tube about 10–15 gm. of dry, elutriated, and graded powder is added; the tubes are then partially filled with bromoform (or other heavy liquid), and the pair or pairs of thimbles are balanced by addition of the appropriate weight of heavy liquid. After centrifuging, a cork stopper that is firmly attached to one end of a brass rod is carefully worked through the layer of light grains near the top of the thimble, and securely plugged into the waist of the tube. The light fraction may now be tipped out without any loss of heavy grains, and the tube above the stopper washed clean with bromoform; the plug is now withdrawn and the heavy fraction poured out, preferably into a small flat-bottomed dish and the tube rinsed out with bromoform. The small amount of bromoform in this is decanted through a small filter into the stock-bottle and the heavy grains washed, and recovered in the manner described earlier. If insufficient residue has been obtained the first time another portion of the original material is poured into the thimbles immediately after decantation of the light fraction, bromoform added again, the tubes balanced and then centrifuged. This process is continued until sufficient heavy residue has accumulated. The upper part of each thimble is then washed clean and the heavy grains recovered in a clean porcelain dish as detailed before. Although this process requires a considerable volume of bromoform most of it is recovered and returned through a filter to the stock-bottle and, by careful manipulation, it is possible to obtain uncontaminated residues with ease.

Similar methods are used for separation of pure mineral samples from rock powders, except that this is not usually a simple separation into two fractions. The rock powder must be carefully sized, and to ensure an absence of composite particles, grains with inclusions, and so on, it is usually necessary to reduce the rock to particles that will pass through bolting silk with a mesh of about 0.125 mm., although finer grinding may be necessary in some instances. Following this the powder should be carefully washed free of dust and dried at 110°C.[5] A plan of order of separation is then worked out so that the minimum number of manipulations will be necessary. As an example let us consider the case of a schist composed of albite, quartz, clinozoisite, spessartite-almandine, prochlorite, actinolite, with minor amounts of iron oxides, titanite, and apatite; a pure sample of prochlorite having a density about 2.96 is desired. In this case one might reasonably proceed as follows: Centrifuge the prepared rock powder in bromoform, portions at a time, until the lower bulbs are filled with material of density

[4] The radius of the thimble-holders is approximately 70 mm.

[5] This temperature cannot be rigidly adhered to since some minerals must be dried at approximately room temperatures otherwise water may be lost. Among the silicates this is true for minerals of the stilpnomelane group and the zeolites. The zeolites are very sensitive to rise in temperature and the usual temperatures employed for mounting rock chips on glass slides with Canada balsam are generally sufficient to change the optical properties significantly. An excellent example of this was observed with a specimen of potassic thomsonite (Hutton, 1942, p. 181) the optical orientation of which was found to be different in crystals mounted in Canada balsam and cooked at the usual temperatures, and those that were merely immersed in refractive index liquids without prior heating.

greater than 2.88 (approximately). Clean the upper parts of the tubes above the stoppers, withdraw the stoppers, and partly fill the tubes with a methylene iodide-bromoform mixture of density 2.98. The small amount of bromoform saturating the mineral powder in the lower bulbs will reduce the overall density very slightly. Shake the tubes and thoroughly mix the mineral powder and liquid; then centrifuge. If the density of the liquid has been adjusted satisfactorily prochlorite will be floating and almost free from any of the denser grains which will have remained behind in the bottom bulbs. To each tube add two or three drops of alcohol and centrifuge once again, and repeat this process until it is evident that the prochlorite would precipitate with any further dilution. At this stage, the required mineral should be absolutely free of any of the denser minerals and in addition it should be freed from any flakes of oxidized or weathered chlorite. Insert the stoppers and decant the prochlorite concentrate into clean thimbles together with the liquid floating it. Add two or three drops of alcohol to each thimble, shake, and centrifuge, repeating this process until the chlorite sinks. Carefully check the surface of the liquid in each thimble for the presence of any grains of quartz or feldspar, and if present insert the stoppers and decant; re-centrifuge the chlorite with the same liquid once again to make doubly certain that no light minerals are occluded with the flakes of chlorite. The purity of the material so prepared may be checked by immersion in a liquid with refractive index equal to that of the chlorite and examined under high magnification. One important requirement is the careful adjustment of the density of the liquid so that the nearly purified powder sinks or rises slowly during the period of centrifuging. For example, if the density of the liquid is very much less than that of the nearly pure powder, the latter will sink rapidly and in a dense clot, and as a result will carry down with it a small but significant percentage of some lighter material. Finally, while on this subject, the author would make a plea for greater use of the centrifuge as a tool for the separation of pure mineral powders from rocks for analytical purposes. Our literature is cluttered up with analyses of impure minerals, because of the universal occurrence of inclusions in minerals. Centrifugal separation is a very sensitive process and with reasonable manipulative skill and care the complete separation of two minerals whose density differs only by 0.01 can be readily accomplished.

Sample Splitting

The total volume of a completely disintegrated specimen is often in excess of the amount required for heavy mineral separation and therefore subdivision is necessary. Unless this is performed carefully, considerable error is likely. Sample splitting has been conveniently accomplished in the past by pouring the powder into a glass filter funnel which is mounted vertically and symmetrically over a divider. The latter is merely an inverted V-shaped piece of brass with a knife-edge ridge; the lower sides of the divider are shaped so that the partitioned powder will pour into two porcelain containers. Such a piece of equipment is easily made in the laboratory and appears to be as efficient as the Jones ore sample splitter.

Since thoroughly representative fractions of heavy mineral residues are required for permanent mounting, electromagnetic fractionation, flotation in other liquids etc., careful subdivision is of the utmost importance and one cannot condemn too heartily the common practice of taking samples with the point of a pen-knife. Heavy residues are generally small in volume, often less than 1000 grains, and consequently it is doubly important that splitting should be as accurate as possible, and the personal factor eliminated. The most satisfactory method involves the use of Otto's microsplit. A second method when a microsplit is not available, requires: a small filter funnel about an inch in diameter at the wide end, with the tube-end ground off perpendicular to the axis of the funnel, a sheet of glass 6″ x 6″, and a razor blade. The funnel is allowed to stand vertically with the ground-off end on the glass plate, and the heavy grains are poured into it. The funnel is tapped gently to shake down any adhering grains and then lifted vertically and slowly off the plate, allowing the grains to arrange themselves into a small cone, which may be quartered down with the razor blade by drawing each portion aside until the desired volume or num-

ber of grains is obtained. It is more accurate to use a funnel to obtain the cone than to pour the sample on to the glass directly from the container, since rolling the grains off a surface usually causes decided concentration of particular grains which may not be distributed evenly within the cone.

Concentration of Samples Before Fractionation in Heavy Liquids

Screening.—In a great many instances, the heavy mineral fractions of many sandstones and so-called "mudstones" are entirely limited to grade-sizes with diameters less than 0.125 mm. In such cases it is often profitable to disaggregate a larger initial sample of rock since a considerable percentage of the rock powder will consist of grade-sizes coarser than 0.125 mm. and these can be discarded. The author has made it routine practice to screen disaggregated sandstones through bolting silk of approximately 0.125 mm. mesh, split off a small fraction of the coarser material, and run a trial heavy liquid separation to make certain that it is free from heavy minerals. Satisfactory correlation of sediments has been made utilizing only the heavy fraction present in the minus 0.062 mm. material, even when a small percentage of "heavies" did occur in coarser screenings. This greatly reduces the amount of laboratory preparation.

In a recent study of the heavy mineral assemblages of some beach sands, screening was carried out in the field after it was found that the heavy fraction was almost entirely limited to grains finer than 0.25 mm.; a screen with mesh openings of 0.5 mm. served to reduce the bulk of samples as much as 70–90 per cent. The original sands were dried in the sun and weighed in specimen bags with a hand spring balance and their weight recorded on the labels attached to screened samples. These samples were further reduced in volume in the laboratory where a more careful watch could be kept for the presence of heavy grains.

Panning.—In view of the experiments of Smithson (1930) and Ewing (1931), considerable care should be taken if panning methods are adopted to reduce the volume of samples. However, in some specimens the percentage of heavy minerals is so low that panning is necessary for economy of heavy liquids and time. Panning of sands in the field is often convenient to reduce the bulk of samples, and in this event, Rastall's (1923, p. 33) advice "to pan a large amount in several instalments not carrying the process too far at first, then adding together all the partial concentrates for a final panning carried out with special care" is worth while.

The reduction of soils by panning to material suitable for ultimate heavy mineral separation in the laboratory is an invaluable aid towards mapping the areal extent of a number of rock-types when a few exposures are available.

Electromagnetic fractionation.—Two types of electromagnets have been devised, one using A. C. and the other D. C. The former device consists of two pole-pieces made from a strip of soft iron, approximately 5″ x 1″ x $\frac{3}{16}$″, bent to a U-shape with a slight closure so that the ends of the pole pieces are about $\frac{1}{4}$″ apart. Two bobbins are wound with 26-gauge silk-covered wire to fit each pole piece, leaving about $\frac{1}{4}$″ of the ends of the poles uncovered; these pole pieces are attached to one end of a wooden handle about 12″ long and $\frac{3}{4}$″ in diameter. On one side of the handle, *viz.* the left side when the instrument is held with the pole pieces pointing vertically downwards, a simple make-and-break switch is attached so that slight pressure from the thumb closes the switch and release of pressure breaks the circuit. This hand electromagnet is connected by 3 or 4 feet of light twin-flex to a rheostat which provides a range of voltage; in actual practice a range of from 2–6 volts has been found most satisfactory.

This instrument is invaluable for clean, rapid separation of magnetite, titanomagnetite, pyrrhotite, odd fragments of steel etc. from samples. The sample is spread out thinly on a shallow aluminum tray, and the magnet, with the switch closed, is moved steadily over the grains at a distance from them that will just allow attraction of the ferromagnetic material but will cause the somewhat less susceptible material such as ilmenite to agitate vigorously; the attracted material is quickly carried aside and dropped in a porcelain evaporating dish by a movement of the thumb on the switch, and the process repeated until the sample is free of ferromagnetic grains. The use of alternating

current instead of direct current has three distinct advantages: the rapid agitation of the grains makes for a clean separation, and since the pole-pieces are completely de-magnetised the moment the circuit is broken, no ferromagnetic grains remain attached thereto; finally the ferromagnetic grains are not themselves magnetized in the process and hence the operator may employ them immediately for hand-picking, screening, and so on, without the difficulty of coalescence of grains.

This magnet is always used to remove magnetite and other ferromagnetic minerals from heavy residues prepared from sediments, blacksands, dredge concentrates etc. before making permanent mounts, or carrying out further electromagnetic fractionation in strong fields.

The second electromagnet uses direct current ranging from 2–12 volts, and is employed for careful fractionation of paramagnetic material. It is an improvement on the instrument designed by Ksanda (1926): (a) The adjustable pole pieces are mounted on the top of the coils, so that manipulation and illumination of the material under examination is much more readily accomplished. (b) The specimen is held in a circular aluminum tray on the underside of which is a pin that enables the tray to be held on an adjustable arm that is attached to one of the corner posts by a well-fitting ring; the height of this arm may be altered by means of the collar and thumb-screw on the post below the arm ring. (c) The post has a fine scale marked thereon to enable the tray to be set at known distances from the apices of the pole pieces.

This instrument has been found invaluable for rapidly obtaining a sufficiency of some rare constituent from a heavy mineral assemblage, for determination of optical properties, chemical tests etc. However, certain precautions must be observed if any electromagnet is to give satisfactory results, particularly when preparing pure fractions: (1) The mineral grains must be washed free from dust. (2) The material for separation should be very carefully graded or, otherwise, a disproportionate percentage of the larger grains in a particular grade-size will be attracted. (3) The mineral grains must be absolutely dry; material for fractionation should be kept in an oven at 90°C. until it is required. (4) The differences between the susceptibilities of ferromagnetic and paramagnetic minerals are so great that before fractionation is attempted, ferromagnetic material should always be removed with the A. C. hand-magnet. (5) Material with average grain diameter of 0.040 mm. can be conveniently fractionated with the large magnet, provided the precautions listed above are observed. With decrease of grain-size, however, obtaining clean separations becomes more difficult owing to mutual adherence of the grains, due partly to adsorbed moisture. To some extent this condition may be overcome by immersing the apices of the pole pieces and the mineral preparation in the pan in either alcohol or ether. (6) It has been found that a D. C. electromagnet of the type described here retains an appreciable amount of magnetism—permanent magnetism—after the circuit is broken, thus preventing free discharge of attracted grains. This circumstance has been overcome by the use of a double-pole double-throw switch for making and breaking the circuit, wired to provide for a momentary reversal of current, after a period of magnetisation, so as to destroy residual permanent magnetism. This switch may be mounted on the bench in front of the operator or more conveniently a simple mercury switch may be mounted on the floor and worked by the foot, thus leaving both hands free for the separation.

Cardboard method.—The cardboard method is a useful adjunct to screening technique for the separation of micas from rock powders. In many instances some degree of concentration of mica can be achieved initially by screening a disaggregated rock powder through appropriately sized screens; careful pounding in a mortar of mica-rich portions so obtained will often reduce the sizes of fragments of brittle minerals but leave the mica plates relatively unaffected. Thus by repeated screening and pounding, samples with fairly high percentages of mica may be obtained. Subsequently the mica in these samples can be nearly completely segregated by treatment on an inclined cardboard tray, consisting of a sheet of fairly smooth cardboard such as Bristol board, about 18″ x 8″ with the longer edges turned up at angles of 90°, glued on a piece of thin plywood or masonite of similar size but leaving about ½″

of the Bristol board unsupported at one end. The tray should be inclined at a low angle, the mica-rich powder spread out thinly near the higher end, and the tray tapped gently. As tapping is continued, the more equidimensional grains will tend to roll or move rapidly down the inclined plane whereas the flakes of mica will lag behind; by careful adjustment of the angle of tilt of the board and the vigor of tapping a high degree of concentration of mica may be obtained. In some instances it is often helpful to breathe heavily over the surface of the board before attempting to separate a powder because the film of moisture so produced seems to obstruct free movement of the mica plates.

If partial concentration of micas has not been attempted by screening as a preliminary measure, then the micaceous rock powders must be reasonably well sized if manipulation of the board is to be in any way efficient.

Hand-Picking and Isolation of Grains

A number of special instruments have been devised for assisting isolation of particular mineral grains from a sample (Howard, 1932; Calkins, 1934 *et al.*) but these have not been found to offer any advantages over the use of simple dissecting needles, fine pointed tweezers, and fine pipettes. For inspection under a binocular microscope it has been found useful to spread the sample out thinly on a smooth card, half of which is black and the other half white. Grains with diameters larger than 0.25 mm. can be removed with the tweezers; for particles of smaller size, a dissecting needle, the point of which has been dipped in a tiny spot (1.0 mm. in diameter) of clove oil or other suitable refractive index oil on a glass slide, is adequate. Grains can be conveniently collected in the hollows of black or white colored spot-test tiles. In order to remove a particle from the point of the dissecting needle and guard against its loss on account of the resilience of the needle, the point of the latter need only be dipped into a spot of clove oil or other suitable liquid on a glass slide or in one of the depressions of the tile. It is often particularly convenient to remove a grain and place it directly in refractive index liquids for immediate observation. Further, if the particle is opaque and less than 0.25 mm. in diameter it may be crushed directly on the glass slide by gently applied pressure from a small agate pestle, and at the same time the liquid serves to prevent loss of any fragments of the crushed grain. This technique was found of considerable value for rapid diagnosis of chromian spinels, hematite, martite etc. since these minerals are normally opaque except in the thinnest fragments.

If it is necessary to remove a particular grain from a mineral assemblage already mounted in oils, the tweezers will again serve for the coarser material. For fine particles, however, the required grain may be safely and conveniently removed from its associates, together with a minute amount of immersion liquid, with a tiny pipette (Lindberg, 1944) and transferred directly to a second glass slide, or alternatively placed in a drop of xylene in a hollow-ground slide to remove the immersion liquid if that should be necessary. This method is much more convenient than the use of gums, vaseline, or Canada balsam on a needle point.

Bolting Silk Screens

Bolting silk has proved to be most useful for sizing fine materials and is especially valuable for splitting up heavy mineral residues. The value of the silk lies in the fact that contamination of the residues, which are usually of small bulk, is unlikely to occur since the silk can be quickly and completely cleaned after each operation; and further with the 1″ diameter screen a very small quantity of material can be graded without loss. Clothholders of different sizes have been designed to permit large or small quantities of material to be graded through one or a nest of silk screens. These holders, 2″, 3″, and 4″ in diameter, are made from seamless brass tubes, machined at each end so that they fit one into the other with sufficient freedom to allow insertion of a thickness of bolting silk between them. If only one holder is required, a simple brass ring holds the silk tautly. For a bottom pan, suitably sized evaporating basins or simply clean, glazed paper are used.

It might be pertinent at this stage to mention that, in the preparation of microslides of fragments of pure minerals for class-work, it has been found useful to grind up the mineral con-

cerned so that it passes completely through bolting silk with approximately 200 meshes per lineal inch. The fine powder is collected in an evaporating dish, washed to remove fine material and dust, and then dried on a hot plate; the material remaining is then of convenient size for study purposes, and the entire operation can be carried through in a very short time.

Diagnosis of Scheelite

Small grains of scheelite are conveniently distinguished from heavy non-tungstenian grains of similar appearance by boiling them in aqua regia in a porcelain dish for 2–3 minutes and subsequently examining them in oblique illumination under a binocular microscope. The characteristic bright-yellow coating of WO_3, readily soluble in aqueous ammonia will have formed on any grains of scheelite.

To determine the distribution of scheelite in a large specimen of quartz an alternative method is preferable. A reasonably large surface should be roughly polished, the specimen placed in a dish together with shot-sized grains of ferrous sulphide, and concentrated HCl added. The vigorous reaction that ensues is allowed to continue for about 3–4 minutes in a fume cupboard, when the specimen is washed vigorously in running water and examined. Any scheelite present will have changed to a grayish-blue color as a result of formation of one of the lower oxides of tungsten, and will accordingly be prominently differentiated from the white or creamy colored gangue. Ferrous sulphide is much more efficaceous than zinc for this test.

Diagnosis of Cassiterite

Cassiterite exhibits a wide range of color from the colorless stream tin of Potosi to the deep blood-red, pleochroic, tantalian variety of Greenbushes, W. A., Westland, N. Z. and other localities. Excluding accurate but time-consuming refractive index determinations, the most simple yet rapid way of proving whether a mineral particle is cassiterite or not, is the reduction method with zinc and HCl. If the grains are numerous and large, it is only necessary to place them in a small porcelain crucible (5 ml. in capacity) together with small shot-sized granules of zinc and add concentrated HCl. However, for a few small (<0.25 mm.) isolated grains the same test can be accomplished without loss of material by the following method: Select a granule of zinc about 4–5 mm. in diameter and punch a depression in it having a volume of approximately 1–2 ml. Next place the zinc granule on a small watch-glass and drop the grain to be tested in the concavity; add a drop or two of concentrated HCl. After 3–4 minutes remove the grain by carefully washing it out of the cavity, dry it, and rub it in a fold of a small piece of chamois leather, smooth side inwards. The grain under test should then be examined with a binocular microscope in oblique illumination when the shiny metallic luster of tin will be obvious if the grain is cassiterite.

Refractive Index Liquids

Numerous media have been recommended (Johannsen, 1918, p. 259–265; Winchell, 1928, p. 74–76; Emmons, 1929, p. 482–483; Holmes, 1930, p. 119–122; Ford, 1932, p. 236; Rogers, 1937, p. 154; Krumbein and Pettijohn, 1938, p. 381–384; Milner, 1940, p. 182–183; Twenhofel and Tyler, 1941, p. 166–167; Switzer, 1944, p. 389; Wahlstrom, 1945, p. 44–45; *et al.*) for refractive index determinations by the immersion method, and a number of authors have simply adopted the list arranged by Larsen and Berman (1934, p. 12). However, so many of the liquids suggested have high viscosities which make careful matching more tedious than it need be. The writer has found that the following liquids serve very satisfactorily for making up sets of refractive index liquids: 2 (butoxy-ethoxy) ethanol (1.43), 1-bromonaphthalene (1.658), methylene iodide (1.74), and methylene iodide saturated with sulphur and tetra-iodo-ethylene[6] (1.81). The mixtures of the first two compounds are clear, almost colorless, and have low volatility and viscosity. For media of higher refractive index West's (1936) liquids should be added and the entire series from 1.43–2.06 made up in steps of 0.01.

[6] The author would like to point out that in his paper on the determination of xenotime (Hutton, 1947a, pp. 143–4) tetra-iodo-acetylene should be read as tetra-iodo-ethylene.

Use of Clerici Solution

The value of Clerici solution for determination of mineral densities and as a high-gravity liquid for mineral separation is well enough known (Vassar, 1925; Jahns, 1939) to require elaboration here, but a note of warning concerning the use of this liquid in some instances needs to be sounded. Pehrman (1935, p. 54–55) has found that the density of ordinary gypsum increased fairly rapidly when immersed in Clerici solution of high density due, presumably, to the formation of thallium sulphate, and that after an immersion period of 42 hours the density of the mineral was in excess of 4. Holzner (1933) and Gruner (1935, p. 704) have shown that minerals of the stilpnomelane group and glauconite will take thallium into their structures when treated with solutions of thallium salts; and further Gruner (1935) has observed that the density of glauconite increased from 2.81–3.02 after being boiled in thallium nitrate for 4 hours. Admittedly this was most drastic treatment and not comparable to the conditions to which a mineral would be subjected under ordinary circumstances, but significant alterations in the densities of minerals may occur within the time required to match the density of Clerici solution with that of a mineral grain, or particularly the time spent separating a pure sample of a mineral from a rock powder. For example, a fragment of chabazite weighing 8.6 mgm. had its density of 2.11, as determined with the Berman balance, increased to 2.21 after 35 minutes in Clerici solution.[7] Further, a tendency for the density of montmorillonite to increase during centrifugal separation in Clerici solution has been noticed by the writer. Therefore it cannot be held that Clerici solution is entirely "stable and inert toward all water-insoluble minerals at ordinary temperatures" (Jahns, 1939, p. 117), and it is deemed advisable to avoid using Clerici solution in work involving minerals showing base exchange properties.[8] This will mean no hardship since such minerals generally have densities less than 3.2 and hence methylene iodide-bromoform and bromoform-acetone mixtures may be substituted for Clerici solution with no inconvenience.

Clerici solution is an expensive material to use, and on those occasions when a liquid of lower density will suffice, it has been found convenient to use the cheaper thallium formate alone, for at a temperature of approximately 30°C. a saturated solution of this salt has a density of approximately 3.60–3.65. Since the solubility of the formate increases very rapidly with the rise of temperature (Sullivan, 1927, p. 21, fig. 10) it is possible to use formate solutions for separations at much higher densities if the operations are carried out in some type of simple calorimeter.

Preparation of Blacksands, Dredge, and Panned Concentrates for Mineralogical Investigation

These samples have high percentages of heavy minerals, sometimes as high as 95 per cent, but rarely less than 40 per cent; for this reason the methods adopted differ considerably from those used for the examination of heavy residues from normal sandstones, siltstones etc. Since no two samples are ever quite alike, some variation of the procedure is often necessary. A brief preliminary examination with hand-lens and microscope, and perhaps some screening, should always be made in order to devise the best procedure.

1. A portion of the original sample, about 75–100 gms., is weighed and washed. If the mineral grains are heavily coated with iron oxide as is often the case with dune-sands, blacksands from deep leads, and dredge concentrates,[9] the iron

[7] Zeolites with very high thallium content have been prepared by Hey and described by him in a number of papers (1932, etc.) but their nature has little bearing on the subject in hand since they were prepared by drastic dry fusion methods.

[8] A similar objection may be raised to the use of Thoulet's solution (a potassium mercuric iodide and also known as Sonstadt's solution) since Hillebrand (1913, p. 439) observed that during use of this liquid for the separation of tyuyamunite (a type of calcium-bearing carnotite) most of the calcium therein was replaced by potassium with profound changes in the optic axial angle. And further Walker (1922, p. 101) found that gmelinite with 0.69 per cent K_2O and 10.08 per cent Na_2O when suspended in Sonstadt's solution for 20 minutes underwent a profound chemical and physical change; an analysis of this zeolite after such treatment showed 14.86 per cent K_2O and only 1.20 per cent Na_2O. Similar results were obtained with okenite.

[9] Fragments of steel from buckets, screens, jigs, etc. often contaminate dredge concentrates and, unless the latter are dried efficiently and quickly before samples are bagged, the minerals therein may be found later to be extensively coated with rust.

stain should be removed before screening, and for this the ammonium sulphide method of Drosdoff and Truog (1935) is most convenient, because little or no solution of any of the constituent minerals will occur.

2. The sample is screened through a nest of wire sieves with openings of 4, 2, 1, and 0.5 mm. in diameter; any material finer than 0.5 mm. is best split up with bolting silk screens, although standard wire screens are also used.

3. Each grade-size is weighed after which ferro-magnetic minerals are removed with the A. C. hand-magnet.

4. The ferro-magnetic fraction of each grade-size is weighed.

5. Each fraction of para-magnetic and non-magnetic minerals is separated in bromoform by the open dish method; generally the quantities of minerals with densities <2.85 are small and the different grade-sizes of these are placed in tubes, labelled, and reserved for later study if that is required.

6. The minerals[10] retained on screen numbers[11] 5 and 10 are hand-sorted, and for this purpose spot-test tiles with 12–16 hollows are very useful in which to collect the different varieties. Sorting according to color has been found most profitable, although subdivision of this coarse material of the basis of density or magnetic susceptibility is sometimes more helpful.

7. For many samples it is advisable to fractionate further a representative portion of each of the $+18$, $+35$, $+60$, $+120$, $+230$, and -230 grade-sizes into fractions with densities <3.85 and >3.85 using Clerici solution.

8. Each of the fractions prepared in the sixth step are now split into paramagnetic and non-magnetic portions; often it is useful to split up the paramagnetic assemblages into samples of different magnetic susceptibility.

9. Precise identification of the minerals in the assemblages of the different fractions is now undertaken.

10. Permanent mounts in both hyrax, aroclor 4465, and Canada balsam are made of all samples of the three finest grades but it is generally of little use to make micro-preparations of coarser materials; these should be preserved in small glass tubes for future reference.

[10] Hereafter the minerals discussed are those with densities > 2.85.

[11] U. S. sieve series.

Grain Counting and Determination of Mineral Frequencies

The relative proportions of the different heavy detrital minerals were determined by counting carefully sized preparations. The latter precaution is most necessary since, if ungraded material is used and the range of grain-sizes is considerable, the sizes of these grains must certainly be taken into account (Stuart, 1927). To enable counting to be carried out quickly and yet in a representative manner a mechanical stage was mounted on the microscope; the stage was racked slowly from left to right and only those grains that touched the intersection of the cross-hairs were counted. At the end of the lateral run the stage was racked an appropriate distance upwards or downwards as the case warranted, and another lateral run was commenced, this time from right to left. If the grains to be counted are fairly large, *viz.* -60 $+120$, it is expedient to count all the grains as they touch the vertical cross-hair during the lateral runs. In every example studied at least 250 grains were counted, and usually 400 were dealt with in this manner. The latter number appeared to be quite adequate since several counts at different times of the same preparation did not show a maximum variation in excess of 5 per cent for the more plentiful constituents. Counting of an additional number of grains uses up a great deal of time and does not contribute importantly to greater accuracy as Dryden's (1931) curves clearly show.

There is no general agreement as to method of representing mineral frequencies. In the past the writer has employed straight percentages of the total number of grains counted, or represented the minerals present as predominant (p), abundant (a), fairly abundant (fa), minor (m), or rare (r). Actual percentages of minerals are not readily compared in some instances, and the inherent difficulties of sampling one's material does not warrant the accuracy implied by strict percentages. The second method of designating the proportions present by names has considerable use in those cases where large

grains are being counted or where, owing to the percentage of opaque constituents, identification of some grains or crystals is open to doubt. The use of names alone without any specification of approximate percentage range makes interpretation by other workers difficult. For instance, what one observer might consider to be a fairly abundant constituent another might believe to be abundant or even predominant. Unfortunately, several schemes involving the use of symbols are in use, such as those suggested by Watts (Milner, 1940, p. 457), Salmojraghi (Boswell, 1923, p. 226) and others, and assessment of these different terms for correlative purposes is not often completely satisfactory.

Therefore, in order to represent the relative abundance of the different minerals present in the three finest grade-sizes, for in these fractions fairly accurate and complete mineral identification was possible, the method used by the geologists of the Burmah and Assam Oil Companies is employed (Evans, Hayman, and Majeed, 1934, p. 39 *et seq.*) in this study. This system overcomes "the difficulties of plotting a truly logarithmic scale by adopting a series of frequency numbers most of which are in geometric progression. This scale is straight forward down to frequency of 3, but for the lower numbers is slightly artificial" (p. 40–41). Their scale is as follows:

Frequency	*Approximate percentage*
8+	90–100 per cent.
8	75–89
8−	60–74
7+	45–59
7	35–44
7−	28–34
6+	23–27
6	18–22
6−	14–17
5	7–13
4	4–6
3	2–3
2	1–2
1	½–1
1*	1 grain only

This system makes comparison between samples simpler, and the minor irregularities in counting due to the personal factor, misidentification of odd minerals, and so on, were reduced in importance; it also has the decided advantage of "weighting" the smaller percentages, and any important differences in the frequencies of minor or uncommon minerals are accentuated.

For grade-sizes of 60-mesh or coarser the following system has been used:

p = predominant (> 60 per cent)
a = abundant (20–59 per cent)
fa = fairly abundant (5–19 per cent)
m = minor constituent (< 4 per cent)
r = rare.

Grain-size of the Dredge Concentrates, Blacksands, Etc.

In general, the mineral assemblages of the dredge concentrates were more complex than the blacksand paystreaks found on open beaches. In some respects this was due to the rarity or absence in most samples from the beaches of material with grain-size finer than 230-mesh; in about 70 per cent of such samples, grains with diameters less than 125 microns were almost absent. In dredge concentrates, on the other hand, material finer than 120-mesh often represented a large proportion of a total sample. This explains, for instance, why uranothorite and biaxial $ThSiO_4$ are only very rare constituents of most of the blacksands of the beaches between Bruce Bay in the south and Saltwater Creek further to the north; the size fraction in which the greater proportion of these minerals are to be found, −120 +230 mesh, is generally absent from such concentrations, but has been carried offshore to where finer particles can settle. Also the conditions for concentration of minerals on beaches is vastly different from those obtaining on the jigs and tables of a dredge, since the latter are adjusted to the optimum position for retention of high-density materials such as gold.

The blacksand paystreaks found along the western coast beaches of South Island generally have rather simple assemblages and contain opaque minerals (chiefly ilmenite) and garnet as the dominant constituents, if some contaminating quartz and feldspars are neglected; they usually show little likelihood of any economic value except for the gold that may be

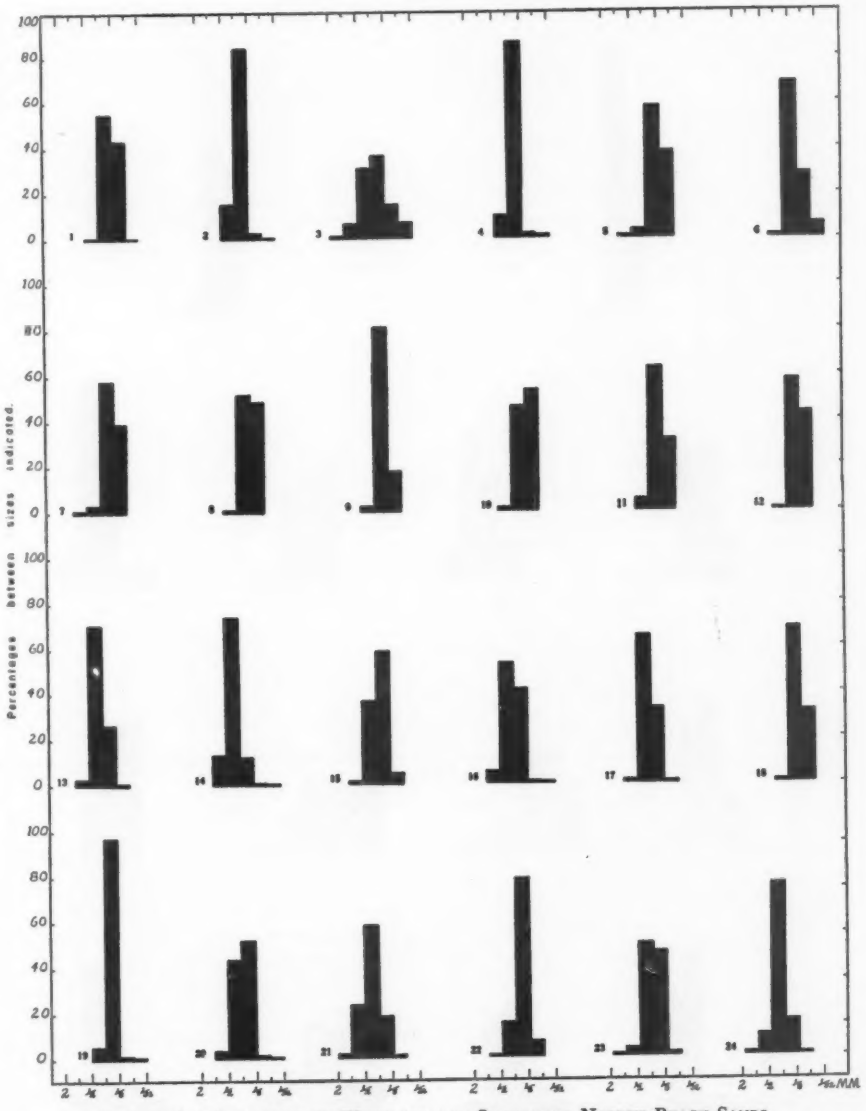


FIGURE. 3—HISTOGRAMS OF WESTLAND AND SOUTHWEST NELSON BEACH SANDS

1. Blacksand, 1½ miles north of Cape Foulwind; 2. Grey sand, 1½ miles north of Cape Foulwind; 3. Mouth of Fox River; 4. Dune sand, Bruce Bay; 5. 1 mile north of Cape Foulwind; 6. Wainui Inlet; 7. West of Charleston Hotel; 8. Karamea Beach; 9. 1 mile south of Buller River mouth; 10. Tauranga Bay, ½ mile south of Cape Foulwind; 11. North end of Nine Mile Beach; 12. Nine Mile Beach, south of Westport; 13. 200 yards north of Five Mile Bluff, Okarito; 14. Five Mile Beach, Okarito; 15. 13 miles north of Greymouth; 16. Ross; 17. 14 miles north of Greymouth; 18. 15 miles north of Barrytown; 19. Paystreak, 14 feet below surface, Gillespie's Beach; 20. 3 miles south of Poerua River; 21. Gillespie's Beach; 22. Hunt's Beach; 23. Bruce Bay; 24. Center of Three Mile Beach, Okarito.

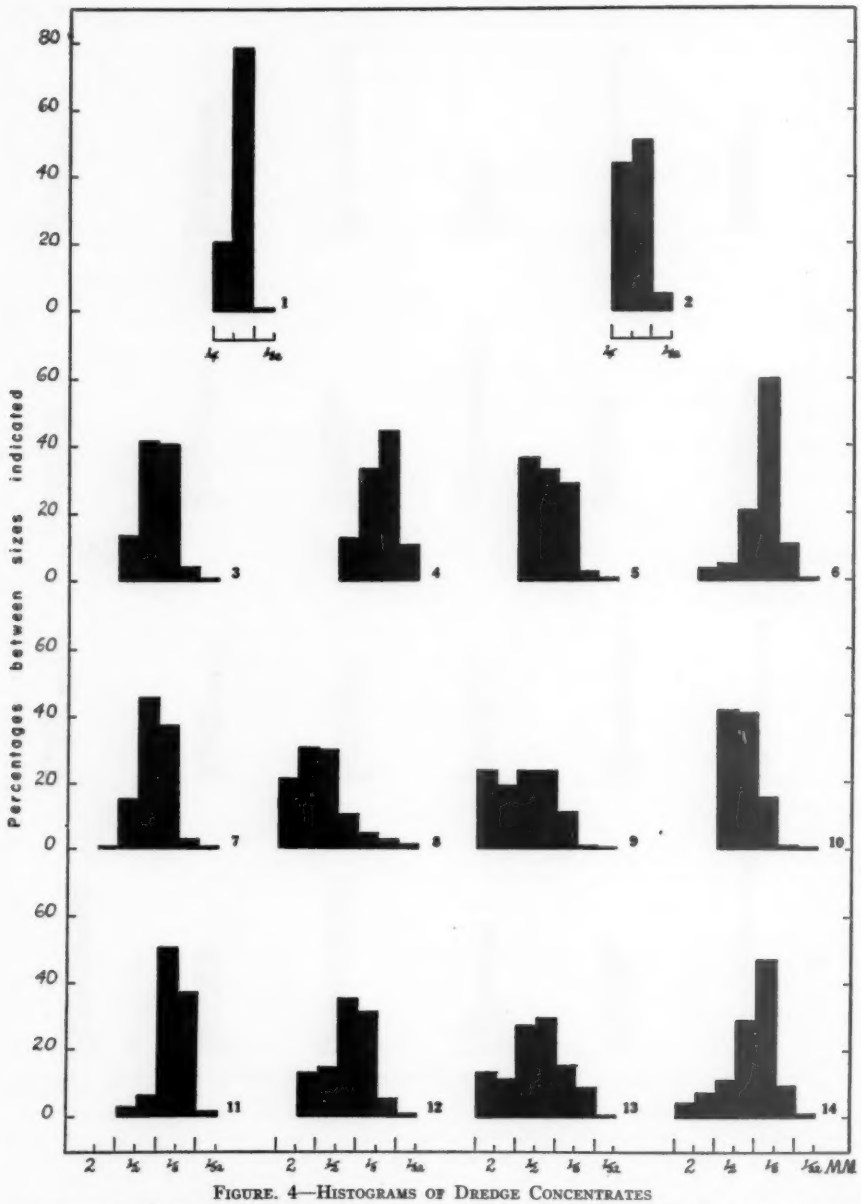


FIGURE. 4—HISTOGRAMS OF DREDGE CONCENTRATES

1. Barrytown; 2. Atarau; 3. Kaniere; 4. Okarito; 5. Lowburn; 6. Ngahere; 7. Red Jacks; 8. Blackball; 9. Rimu; 10. Arahura; 11. Gillespie's Beach; 12. Snowy River; 13. Grey River; 14. Slab Hut.

contained in them. However, the beach sands, dune and terrace materials of the central to southern Westland coastal strip contain interbedded with the mineralogically simple black-sand lenses or paystreaks, the finer-grained silts in which thorites are usually found. Having determined that most minerals of any value were restricted for the most part to the finer grades, size analysis was a rapid means of determining the possible economic value of a beach sand. Size analyses of samples of some of these sediments are illustrated in Figure 3, and it will be observed that most of these sands are well graded since the bulk of the particles are restricted to either one or two size fractions.

The grading of particular dredge concentrates varied to some extent from time to time but the range in any individual case was not excessive. Therefore, for the sake of comparison with the beach sands, grading analyses of at least three samples taken at different times from a number of dredges have been averaged and the results presented in histograms (Fig. 4).

In some instances magnetite and other ferromagnetic material had been removed from samples of dredge concentrates before they were made available for this work but where such material was present when collected or sent to the writer, it was removed electromagnetically before size analyses were made.

Part 2. Mineralogy

The chemical and physical properties of all the minerals in the concentrates, sands, and gravels collected by, or made available to the writer have been determined as completely as possible, and the outcome is detailed in this section of the work. A few grains which defied satisfactory diagnosis, since they are never more than rarities, are listed as unknowns without any qualifications. The occurrences of the mineral species, with their frequencies are set out in Table 19, but every sample studied has not been listed here. For example, of over 30 concentrates available from the Arahura River, three analyses that appeared fairly representative have been listed. Occasionally the averaged analyses of four or five concentrates, obtained from a dredge or sluicing claim over a period of time are listed. Other properties such as solubility, stability, and alteration have been discussed when needed and a few of the minerals which have somewhat unusual properties have been discussed in some detail.

Actinolite: Actinolitic amphiboles have been recorded in concentrates of two dredges, Lowburn Ferry, near Cromwell in Central Otago, and Snowy River, in Southwest Nelson, and from pannings of the sands and gravels of Hokitika River and Kyeburn Stream and its tributaries. Actinolite is most abundant in Lowburn material but does not make up more than 2 per cent and 1 per cent of the −230 and −120 +230 grades respectively; it occurs in short, prismatic crystals with weak but distinct pleochroism: X = very pale yellowish-green; Y = green; Z = bluish-green. The beta refractive index was determined as 1.648 ± 0.003, with $\gamma \wedge c = 17–19°$, and 2V approximately 60°. The amphibole is therefore comparable to the variety frequent in the green schists of Western and Central Otago (Turner, 1933; Hutton, 1940a, p. 43), and is an actinolitic member of the actinolite-glaucophane series as defined by Kunitz (1930). Similar properties were recorded for Hokitika River material. Occasional acicular amphibole fragments, present in −230 mesh fractions of Bruce Bay beach sand concentrates, differed somewhat from the actinolite just described. Here the Z-vibration direction is blue to mauve, the birefringence is lower, about 0.014, and a slight dispersion is evident. The amphibole is similar in some respects to crossite from Coronet Peak (Hutton, 1938b, p. 210–211) but differs from that mineral in that the axial plane is parallel to (010). This Bruce Bay mineral is believed to be an aluminous member of Kunitz (1930) glaucophane-riebeckite series.

Anatase: Anatase (Pl. 2, fig. 33) is a rare constituent of a beach sand (Wainui Islet) as simple pyramidal forms with prominent development of basal plane *c* (001). Striations are well developed parallel to the prism-pyramid join. The color is yellow to yellowish-brown with only faintly perceptible dichroism; tiny clouds of fine dust-like material, presumably magnetite or ilmenite, were the only inclusions noted.

Andalusite: Andalusite was recognized only in the finest fraction (−230 mesh) of Snowy

River material. It formed irregularly terminated, prismatic, unaltered grains, devoid of inclusions and with distinctive pale-pink to pale-green pleochroism. The refractive index β was found to be 1.634 and 1.636 for two fragments.

APATITE: Although apatite is present in most of the samples examined, the percentage in

PLATE 1.—TYPICAL FRAGMENTS AND CRYSTALS OF ZIRCON AND HYACINTH

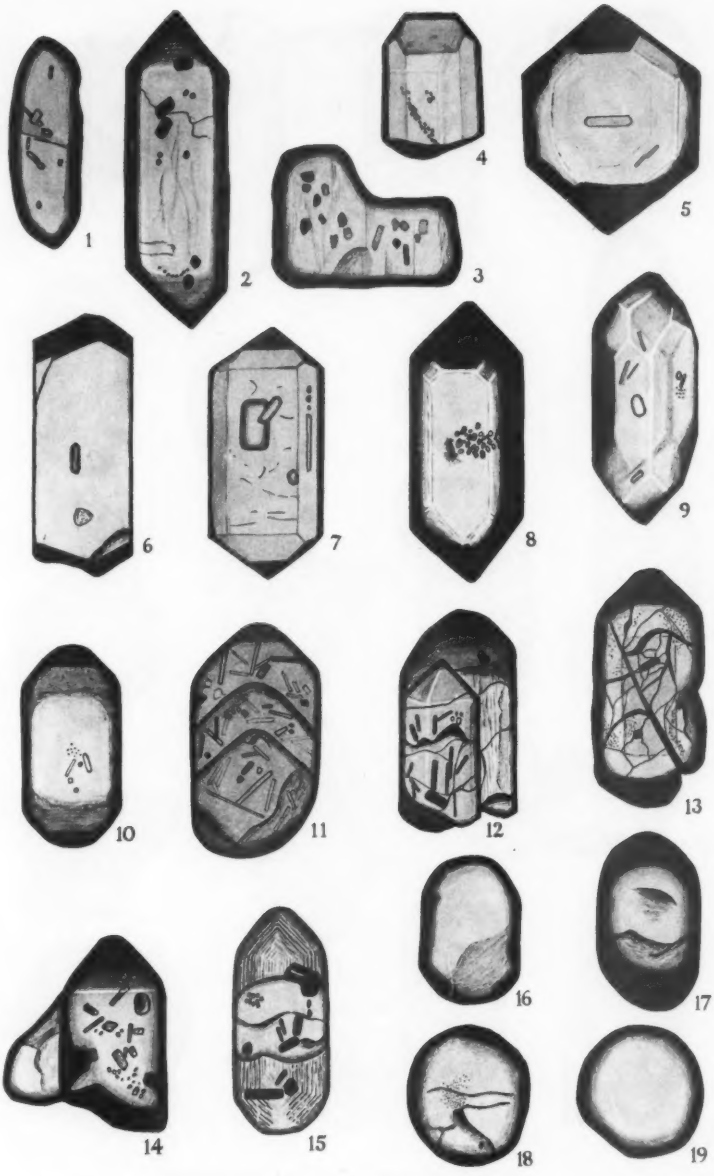
Figure

1. Zircon, showing moderate degree of rounding. Barrytown. ×90.
2. Zircon, with simple development of first order prisms and pyramids. Gillespie's Beach. ×125.
3. Zircon, probably a crystal broken in a rather unusual manner and subsequently subjected to considerable abrasion. Kaniere. ×105.
4. Zircon showing development of (110) and (100) prisms, with (111) pyramids. The (001) face developed here is unusual. Atarau. ×200.
5. Zircon with (110), (100), (111), and (101). Short stumpy habit uncommon. Recent beach sand, Wainui Inlet. ×105.
6. Zircon, common type. Barrytown. ×140.
7. Zircon, simple type, not common. Barrytown. ×105.
8. Zircon with development of (110), (100), (111), and (311). Okarito. ×105.
9. Zircon, common type. Barrytown. ×115.
10. Zircon, purple variety of characteristic simple habit. Ngahere. ×65.
11. Zircon, purple variety with zonary structure. Ngahere. ×65. Mounted in hyrax.
12. Zircon, purple variety including a second sharply euhedra crystal of purple zircon. Atarau. ×105.
13. Simple crystal of purple zircon with faint zoning and showing very extensive fissuring. Ngahere. ×75.
14. Simple purple zircon with outgrowth of similar color. Ngahere. ×43.
15. Purple variety of zircon with delicately developed zonary structure, and a nucleus of intensely abraded purple zircon. Atarau. ×75. Mounted in hyrax.
16. Well rounded purple zircon, devoid of inclusions and without zoning. Ngahere. ×80.
17. Well rounded purple zircon. Ngahere. ×80.
18. Purple zircon, well rounded. Ngahere. ×80.
19. Extremely well rounded purple zircon. Ngahere. ×80.

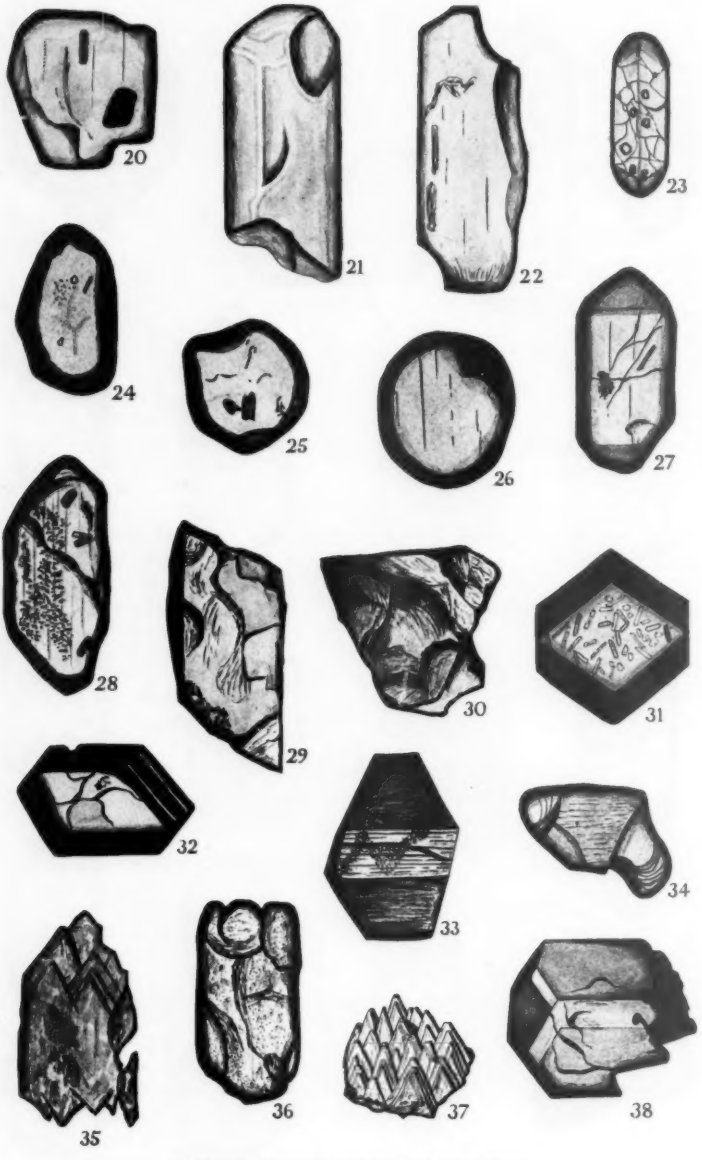
PLATE 2.—TYPICAL HEAVY DETRITAL MINERALS

Figure

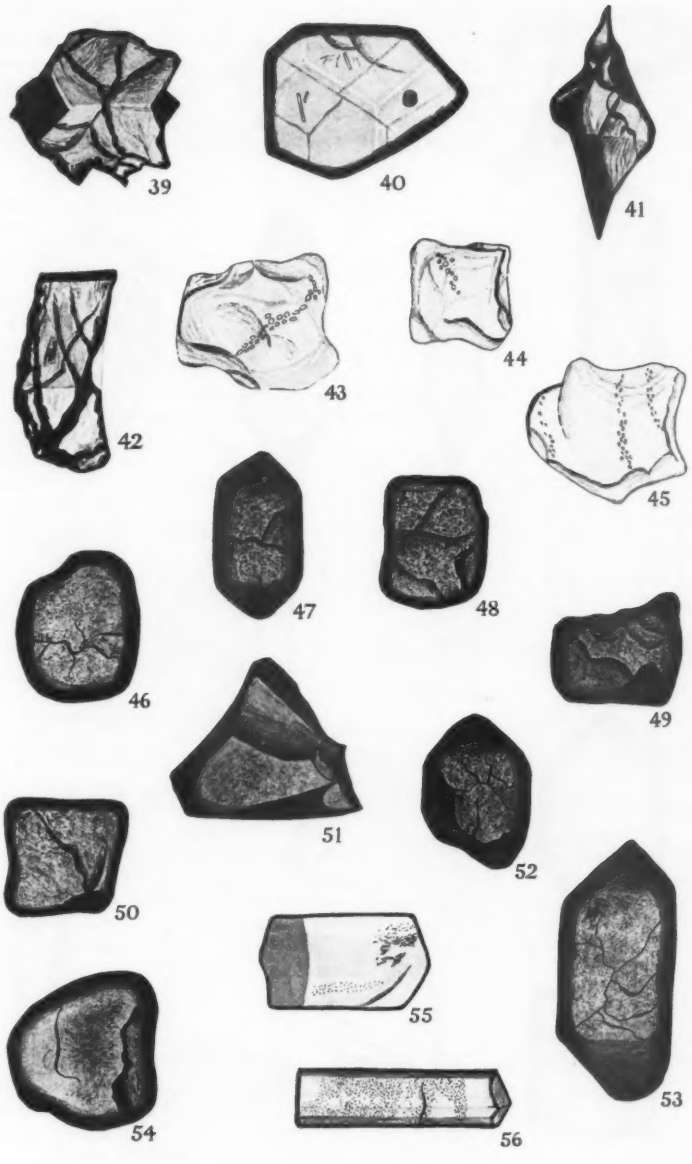
20. Apatite, Arahura. ×120.
21. Apatite, simple crystal and little worn. Arahura. ×145.
22. Apatite. Arahura. ×145.
23. Idioblastic ferriferous epidote with characteristic system of fractures. Lowburn. ×50.
24. Worn grain of almost colorless monazite. Barrytown. ×180.
25. Monazite. Barrytown. ×180.
26. Well rounded monazite. Barrytown. ×125.
27. Euhedral tablet of monazite parallel to (100) showing poor development of (010) cleavages. Barrytown. ×125.
28. Monazite, similar to fig. 27. Atarau. ×75.
29. Cassiterite. Barrytown. ×105.
30. Cassiterite. Barrytown. ×105.
31. Sharply euhedral spessartite. Lowburn. ×85.
32. Almandine. Arahura. ×125.
33. Yellowish-brown anatase. Recent beach sand, Wainui Inlet. ×125.
34. Somewhat worn grain of cinnabar, showing perfect cleavage. Barrytown. ×105.
35. Scheelite. Barrytown. ×105.
36. Scheelite. Barrytown. ×105.
37. Fragment of scheelite. Barrytown. ×145.
38. Euhedral titanite. Recent beach sand, Wainui Inlet. ×40.



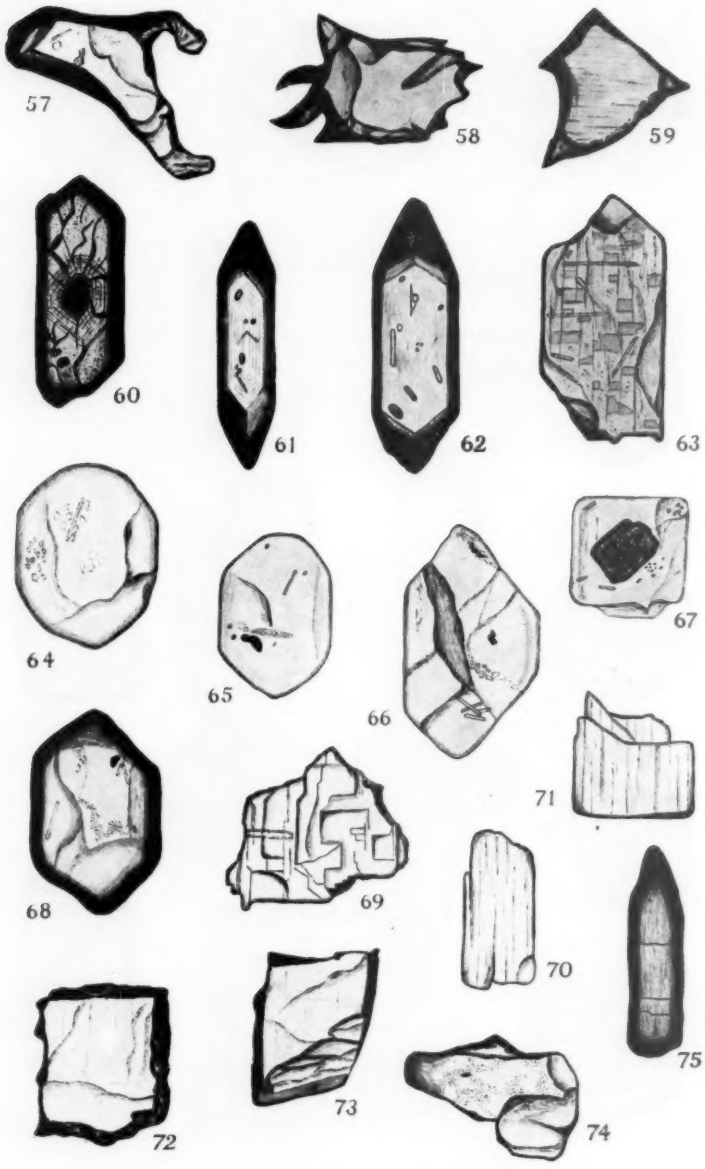
TYPICAL FRAGMENTS AND CRYSTALS OF ZIRCON AND HYACINTH



TYPICAL HEAVY DETRITAL MINERALS



TYPICAL HEAVY DETRITAL MINERALS



TYPICAL HEAVY DETRITAL MINERALS

the different assemblages is never high except in beach sands from the Ligar Bay—Separation Point area in north-west Nelson. It usually occurs as slightly worn, uni-terminated, rather stumpy, prismatic crystals (Pl. 2, figs. 21–22), although well worn or irregularly shaped anhedra (Pl. 2, fig. 20) are usually present, and dominant in some cases. The grains of apatite

PLATE 3.—TYPICAL HEAVY DETRITAL MINERALS

Figure

39. Titanite. Wainui Inlet. ×30.
40. Titanite. Wainui Inlet. ×150.
41. Titanite. Wainui Inlet. ×60.
42. Titanite. Wainui Inlet. ×35.
43. Topaz. Okarito. ×235.
44. Topaz. Okarito. ×165.
45. Topaz. Okarito. ×135.
46. Worn grain of uranothorite. Gillespie's Beach. ×80.
47. Uranothorite, showing simple development of (110) and (111). Gillespie's Beach. ×90.
48. Uranothorite, common rectangular type. Gillespie's Beach. ×70.
49. Uranothorite. Gillespie's Beach. ×60.
50. Uranothorite, common rectangular type. Gillespie's Beach. ×60.
51. Uranothorite. Gillespie's Beach. ×75.
52. Uranothorite. Harihari. ×60.
53. Sharply euhedral grain of uranothorite. Harihari. ×115.
54. Uranothorite. Harihari. ×75.
55. Tourmaline, clear type. Barrytown. ×180.
56. Tourmaline, typical "schist" type. Lowburn. ×40.

PLATE 4.—TYPICAL HEAVY DETRITAL MINERALS

Figure

57. Strongly etched garnet. Rimu Dredge. ×105.
58. Strongly etched garnet. Snowy River Dredge. ×125.
59. Sub-euhedral table of β-zoisite, oriented $\perp$ to Bxa, and showing (010) cleavage. Grey River Dredge. ×195.
60. Hyacinth, showing fine zonary structure surrounding a nucleus of opaque material. Slab Hut Dredge. ×175.
61. Colorless zircon showing prominent development of steep pyramid (331) and very fine zonary banding. Snowy River Dredge. ×100.
62. Colorless euhedral zircon showing development of steep pyramid (331). Snowy River Dredge. ×100.
63. Hypersthene with pale brown inclusions (schiller-structure). Snowy River Dredge. ×215.
64. Moderately rounded euhedron of xenotime; embedding material is hyrax (n = 1.70 approx.). Snowy River Dredge. ×215.
65. Same description as figure 64.
66. Same description as figure 64.
67. Slightly worn grain of xenotime oriented with Z perpendicular to the plate; such grains give centered uniaxial interference figures. Mounting medium is hyrax. Large inclusions of brown unidentified mineral with high refractive index. Snowy River Dredge. ×110.
68. Slightly worn grain of xenotime mounted in Canada balsam. Snowy River Dredge. ×215.
69. Irregularly terminated bladed crystal of kyanite. Red Jacks Dredge. ×180.
70. Slender, flattened prism of sillimanite. Rimu Dredge. ×140.
71. Stumpy prism of sillimanite. Snowy River Dredge. ×140.
72. Colorless (001) cleavage fragment of monazite showing traces of cleavages parallel to (100) and (010). Blackball Creek Dredge. ×240.
73. Same description as figure 72.
74. Irregular grain of staurolite with marked conchoidal fracture. Blackball Creek Dredge. ×105.
75. Deep red prismatic grain of rutile with rounded terminations. Blackball Creek Dredge. ×105.

are usually colorless and water-clear with a complete absence of dichroism. Inclusions are not distinctive but magnetite, titanite, and zircon have been recognized; strings of minute cavities have been observed in some grains, with swarms of indeterminate needles, oriented parallel to the *c* crystallographic axis, in others. In the Ligar Bay beach sands apatite is such an important constituent that it was possible to study it in greater detail than in other occurrences. In this material the apatite occurs in stout prismatic crystals with a ratio breadth: length of approximately 1:2 in the finer gradesizes; in the coarser screenings, viz. −60 +120, because of considerable wave action, the crystals are more equant, with subhedral to anhedral forms predominating. In the finer grades, apatite crystals are usually terminated at both ends by basal plane and a combination of pyramids. An imperfect cleavage, parallel to (0001) can be observed in most grains, and usually the shape of fractured grains has been determined by this incipient cleavage direction. Inclusions are fairly common and more interesting in this material; the following details have been noted: (1). Tiny grains of what seem to be sharp euhedra of apatite. (2). Lines of brown needle-like bodies, arranged in bands, both bands and needles oriented parallel to the *c* crystallographic axis. (3). Patches or clouds of dust-like unidentifiable material that gives the grains a smoky aspect when viewed under low magnification. (4). As in (2) above, but whereas the needles are arranged parallel to the *c* axis the bands are perpendicular thereto. (5). Rows of tiny globules, possibly gas filled, perpendicular to the *c* direction. Measurements of refractive indices for a number of crystals from the Ligar Bay area showed a range as follows: E = 1.633–1.635 ± 0.001; O = 1.638 − 1.640; O − E = 0.005. The mineral therefore appears to be a normal fluorapatite; microchemical tests showing only a trace of chlorine and an absence of hydroxyl support this conclusion. Further qualitative chemical tests (following the method of Watson, 1935, p. 28) proved that silica was present in crystals devoid of inclusions, but the location of the silicon in the structure has not been determined. It is sufficient to state at this stage that replacement of P^{5+} by Si^{4+} would require a balancing substitution to occur elsewhere in the structure, which could be effected with replacement of Ca^{2+} by Y^{3+} as Wickman (1943, p. 387–8) has suggested. Extensive substitutions of this nature are known as in the mineral britholite (Hägele and Machatschki, 1939) and its isomorph, the yttrium silicate apatite, abukumalite (Hata, 1938a). It would be interesting to determine whether the distinct range of hardness found in different specimens of apatite can be correlated with the $CaP \rightleftarrows YSi$ substitution, since a hardness of 6 has been reported for abukumalite.

BARITE: Very rare, slightly abraded cleavage fragments of barite recognized among the finest fraction of concentrates from Grey River dredge are colorless and devoid of inclusions, the β vibration direction of the refractive index ellipsoid was determined as 1.635–1.637. They appear to be basal sections, their shape determined chiefly by cleavage parallel to 001.

BIOTITE: As expected of a relatively low-density and flaky mineral, biotite is a minor accessory, although it occurs in many of the samples studied. By way of contrast, heavy residues from a great many Tertiary sediments from Westland localities are exceedingly rich in brown mica. Red-brown plates with strong to intense pleochroism and small optic axial angle are most frequent. The mica may be very pale brown (Ligar Bay) or greenish-brown (Arahura River); in the latter some of the mica is oxidized in part to reddish-brown plates, and the mineral appears to become uniaxial or very nearly so.

CASSITERITE: Cassiterite, though recognized in every Westland and Nelson dredging examined, is no more than a mineralogical curiosity except in Ngahere material where it makes up approximately 13 per cent of the −60 +120 screening, and Blackball samples. Generally it occurs in conchoidally fractured anhedra of pronounced irregular form. Grains similar to those illustrated (Pl. 2, figs. 29–30) are typical; rarely traces of crystal faces were observed and twinned crystals are rare.

Most cassiterites are deep red to reddish-brown; pale to colorless examples are much less common. The strongly colored varieties exhibit intense dichroism quite comparable to that characteristic of members of the tapiolite-mossite series. In the preliminary stages of the

study the exact nature of the mineral was in some doubt. Dichroism in most examples followed the scheme: O = pale greenish-yellow. E = deep reddish-brown (blood-red in some instances). Absorption: E > O.

Although a uniaxial positive character was general, biaxial interference figures with small optic axial angles (approximately 10°) were found, the sign, however, remaining positive. The biaxial character of cassiterites that do not show strain effects is not easy to explain, but Gotman's (1939, p. 472) assumption that such anomalies are connected with the development of zonary structures is not supported. Although zoning was observed in a few cases, some unzoned crystals are biaxial.

Refractive index determinations by the immersion method on several varieties of cassiterite gave results as follows:[12]

Clear red	=	1.996
Semi-opaque red	=	1.994
Deep red-brown	=	1.998 2.001 2.008
Zoned cassiterite:		
Red central zone	=	1.997
Pale brown peripheral zone	=	1.993

These values accurate to ±0.003, compare with those generally recorded (Milner, 1940, p. 259; Winchell, 1933, p. 52), although the refractive index of the ordinary ray in one deep red-brown variety is slightly higher than in normal cassiterite.

Density determinations gave a range of 6.72–6.91. A distinctly blotchy effect characterizes much of the material examined; this feature has been commented upon by other writers (Milner, 1940, p. 259).

The intensity of absorption in these cassiterites distinguishes them from those previously described from New Zealand. The cassiterite from the eluvial deposits of the Tin Range, southern Stewart Island (Williams, 1934, p. 352) is opaque or pale amber in thin splinters, an observation confirmed by the writer's examination of several specimens of cassiterite from this and other localities in the vicinity of Port Pegasus. Cassiterite from pegmatites in Doubtful Sound area is colorless (Turner, 1937b, p. 242–243), although some (Hutton and Turner, 1936, p. 259) in Tertiary sandstones from the Lake Manapouri area was deep red-brown but dichroism was not distinctive. However Rastall (1919, p. 272) has recognized cassiterite with red to grayish-brown dichroism in heavy residues from the Lower Greensand of Eastern England, whereas Scrivenor (1919, p. 123–124) recorded cassiterite from Siam with an intense dichroism, E = deep red, O = green, which he believed was due to the presence of titanium, although the analysis showed only 0.17 per cent of TiO_2 to be present. Groves (1932, p. 210) has described very pale or nearly colorless cassiterite with deep blood-red zones, also unzoned varieties with intense dichroism, brownish-green to deep blood-red (1932, p. 207), and has found blood-red cassiterite in the detritus from the Dartmoor granites (1931, p. 74). Among the minerals of the Center Strafford pegmatite, Switzer (1938, p. 817) found cassiterite with colorless to dark reddish-brown dichroism, and Stheeman (1932, p. 144) noted a pink tint in zonal cassiterite from Uganda.

The intensity of absorption of some of the Westland cassiterite and its physical similarity to tapiolite suggested the presence of tantalum; certainly the ionic radii of tantalum (Ta^{4+} = 0.68 Å.) and tin (Sn^{4+} = 0.74 Å.) are close enough to allow some substitution without any rupture of the structure. With this in view, a portion of the concentrates from Ngahere Dredge was treated by screening and electromagnetic methods and in a sample containing a high percentage of cassiterite and somewhat contaminated with zircon, Mr. Seelye found 2.3 per cent of Ta_2O_5[13]; thus tantalian cassiterite (ainalite of A. E. Nordenskiöld or tantalocassiterite of Schoeller, 1937, p. 31) would appear to be an appropriate term for the red Westland cassiterite. Tantalum in cassiterite[14]

[12] The values for N_o only were determined as accurately as possible; N_e was not found but is in excess of 2.07 in the grains employed.

[13] The absence of niobium may be unusual since, although niobium is found in the second and tantalum in the third long periods, their ionic dimensions are almost identical as a result of the lanthanide contraction, and their presence together in cassiterite might be expected.

[14] While this work was in manuscript form, Liebenberg (1946) has shown that the dichroism of cassiterites from S. W. Africa and S. Rhodesia is due to small quantities of tantalum and niobium in solid solution.

is probably far more common than it appears to be; Headden (1906, p. 167), Simpson (1910, p. 310), Schaller (1912), p. 18–22), and others have observed it, although analyses report not more than about 5 per cent of Ta_2O_5[15], whereas the extent of the replacement of tantalum by tin in tapiolite, mossite, or tantalite-columbite appears to be about 2.5 per cent[16]. Schaller (1912, p. 22) was of the opinion that miscibility of cassiterite and tapiolite was complete, but from a structural point of view there does not appear to be room for extensive substitution because some other ion, probably a trivalent one such as Fe^{3+} would have to be introduced into the lattice for compensatory purposes.[17] Laboratory experiments by Quensel (1941) have suggested that cassiterite may take up about 25 per cent of tapiolite into solid solution, but whether the compounds so formed are stable or metastable only at ordinary temperatures is not explained. On the other hand, Edwards (1940), finding that cassiterite with 3.86 per cent of Ta_2O_5 from Greenbushes, Western Australia, contained inclusions of tantalite, concluded that at high temperatures cassiterite and tapiolite (and perhaps mossite) are isomorphous but that exsolution of the greater proportion of one component occurs at ordinary temperatures. No analyses, together with evidence of homogeneity, are available, at least to the present writer, to show the existence of complete ionic substitution between cassiterite and tapiolite, and a recent re-investigation of a number of old specimens of ainalite and ixiolite by Åmark (1941) has proved that they are mixtures of cassiterite and columbite or tapiolite and not truly homogeneous minerals; in addition the X-ray studies of Brandt (1943) have clearly proved that a tin-bearing manganese tantalate from Varutrask, Sweden, is merely a mixture of cassiterite and columbite.

Polished sections of Westland cassiterites containing 2.3 per cent of Ta_2O_5 showed no indication of inhomogeneity resulting from exsolution and it must be assumed, then, that the tantalum is in solid solution. This and the data supplied by Edwards clearly suggest that the limit of solid solution of tantalum in cassiterite lies somewhere between 2.3 and 3.86 per cent of Ta_2O_5.

CHLORITE: Chlorite group minerals are entirely restricted to the finer grade-sizes of the samples studied, and they belong to the prochlorite group as defined by Orcel (1927, p. 341); they compare closely with chlorites in the metamorphic rocks of Western Otago (Hutton, 1940a). For isolated flakes the following data were determined:

Lowburn Dredge
β = *1.633–1.635;* $\pm$ *0.002*
(*in four samples*)
Elongation. = *Z*
Dispersion. = $\rho > \nu$
Anomalous tints: Purple to blue.

Arahura Dredge
β = *1.615–1.620;* $\pm$ *0.002*
(*three samples*)
Elongation = *X*
Dispersion = $\rho < \nu$
Anomalous tints: brown.

The chlorites were typically pale grass-green and the pleochroism was faint but distinct in all cases. In occasional flakes some oxidation had occurred, resulting in (a) a change of color from green to greenish-brown or brown, and (b) a distinct rise in the refractive indices and birefringence (Winchell, 1936, p. 644–645). One oxidized chlorite gave the following data: α = 1.637; γ = 1.654; $\gamma - \alpha$ = 0.017. This change in chlorites results in a product not unlike some members of the stilpnomelane group of minerals, but most of the chlorites may be readily distinguished by their lower refraction (Hutton, 1938a, p. 187), weaker absorption, and absence of brittleness; members of the stilpnomelane group are always optically negative.

CHLORITOID: The only fragments of chloritoid

[15] 2.34% Ta_2O_5 in cassiterite from Kara-su, Turkestan Range, Fioletova, 1940; 4.64% Ta_2O_5 in cassiterite from Niriella, Ceylon, Palache *et al.*, 1944, p. 576; 3.26% Ta_2O_5 and 0.04% Nb_2O_5 in cassiterite from Kimito, S. W. Finland, Pehrman, 1945. (Whether these cassiterites are homogeneous is unknown.)

[16] 1.26% SnO_2 in tapiolite from Skogböle (Schaller, 1912, p. 19); 0.53% SnO_2 in tapiolite from Finland (Pehrman, 1932); apparently in traces only in tapiolite from Belgian Congo (Buttgenbach, 1933); Sn is absent in tapiolite from Cresciana (Casasopra, 1938); 2.16% SnO_2 in tantalite from Brazil, (De Almeida *et al.*, 1944, p. 218).

[17] The electrical balance of the lattice could also be maintained by simultaneous addition of Sc^{3+} (Rankama, 1948, p. 29).

recognized during this investigation were found in the −35 +60 fraction of one of the concentrates from Snowy River dredge. The mineral was sieved with inclusions of quartz, sodic plagioclase, and magnetite, and a value of 1.721 was found for the beta direction of the refractive index ellipsoid; a distinctive pleochroism follows the scheme: X = pale bluish-green; Y = greyish-blue; Z = pale green; $Y \geqslant X > Z$. The birefringence was low, certainly not in excess of 0.01, and strong zoning was apparent. These properties are almost identical with those of chloritoid observed by the writer in slates in the Anatoki Valley, northwest Nelson.

ALUMINIAN CHROMITE: Aluminian chromite is the only member of the spinel group, excluding magnetite, of any importance in the samples studied. It is well represented among some of the coarser screenings, ranging up to 21 per cent in one sample (Blackball dredge), but is rare in finer fractions. Chromite particles range from sharply euhedral octahedra to well-rounded forms in the coarse grade-sizes; fragments with pronounced conchoidal fracture are usual in the finest screenings. The mineral is quite opaque except in the thinnest flakes, but observed in oblique illumination, the worn chromite grains had a strikingly smooth, brilliant, jet-black appearance, and if broken, a very pronounced conchoidal fracture; occasionally octahedra were noted with well developed natural etch marks in the form of deep triangular pits (Ngahere dredge) on the octahedron faces. Comparable etch-pits were produced artifically but with difficulty on chromite octahedra by treatment with perchloric acid (Frankel, 1940, p. 198–199), and treatment of the mineral grains for 5 minutes in molten potassium pyrosulphate had a similar effect. These properties, quite distinct from those shown by the associated opaque constituents, permitted reasonably accurate estimation of the quantities of this constituent. Confirmation of the identity of each grain was obtained by removing the particles on the point of a needle or by means of fine-pointed forceps, crushing them finely in clove oil directly on a glass slide if possible, and examining the powder in transmitted light. A yellowish-brown color is characteristic and in no instance were reddish tints observed.

Sufficient octahedra, or portions thereof, were separated from Slab Hut dredge concentrates in which they are particularly plentiful, by hand-picking beneath a binocular microscope, and quantitatively analyzed (Table 1).

The determination of ferrous oxide in refractory minerals is extremely difficult; several methods were tried by Mr. Seelye but none were entirely satisfactory, although Seil's[20] method gave the best results; however some preliminary experiments with Rowledge's (1934) sodium metafluoborate method gave promising results. Therefore before calculation of the structural formulae, correction of the FeO: Fe_2O_3 ratio was made in order to satisfy the requirements of the spinel formula. Except for the small percentage of SiO_2 in analyses C and E, the type of ionic substitution is quite usual. SiO_2 has been ignored in these calculations because the writer has no definite information as to whether it is present as a mechanically mixed impurity or in solid solution in the spinellids. Careful microscopic examination of the purified and powdered minerals did not reveal any adhering serpentine, talc, or chlorite, and the writer would suggest that the silicon was in solid solution. Silica is not uncommonly recorded in analyses of members of the spinel family, and although it is not always certain that the analysed material was pure, silica has been recorded in very carefully purified spinel by Vincent (Tilley, 1938, p. 83), and strength is added to the suggestion of solid solution by the work of Rankin and Merwin (1918, p. 307) who found that synthetic spinel with about 5 per cent of SiO_2 appeared to be homogeneous.

For comparison with other chromites, the compositions of the New Zealand minerals may be expressed as end-members per unit cell.

Vis.

	B	D	F
Spinel	3.63	4.14	4.13
Magnesiochromite	1.69	1.55	1.66
Ferrochromite	2.01	1.95	1.73
Magnetite	0.67	0.36	0.48

Stevens (1944, p. 31) has conveniently classified the chromium-bearing spinels into six main groups by means of a ternary diagram, and if the analyses of the New Zealand chromites are

[20] J. Ind. Eng. Chem. Anal. Ed. 15 (1943), p. 189,

TABLE 1.—ANALYSES AND PHYSICAL PROPERTIES OF ALUMINIAN CHROMITES

	A	B	C	D	E	F	G	H
Cr_2O_3	36.37	38.29	43.46	43.67	42.80	43.45	35.5	46.16
Al_2O_3	24.72	26.03	24.60	24.72	23.65	24.01	26.2	10.89
Fe_2O_3	—	7.45	1.58	4.07	3.78	5.26	7.9	2.78
FeO	20.47	13.36	13.54	11.36	12.12	10.90	12.4	20.48
MgO	14.12	14.87	15.87	15.93	15.84	16.08	15.2	11.04
TiO_2	1.54	—	0.34	—	0.46	—	0.3	—
SnO_2	2.24	—	—	—	—	—	—	—
MnO	trace	—	0.24	0.25	0.17	0.18	0.1	0.26
SiO_2	—	—	0.38	—	0.40	—	0.2	5.76
H_2O	—	—	—	—	—	—	—	2.21
CaO	—	—	0.23	—	0.20	—	0.1	—
CoO	—	—	—	—	0.12	0.12	—	—
NiO	—	—	—	—	0.04	—	—	—
	99.46	100.00	100.24	100.00	99.58	100.00	99.2	99.58
n^{18}	1.972	—	1.970	—	1.970	—	—	—
D. 20° C.	4.67	—	4.67	—	4.66	—	—	—

A. Spinel from Slab Hut Dredge concentrate, 20 chs. 15° from Trig. S. T., Mawheriti S. D. Owing the small amount of the sample available and its extremely refractory nature, ratio $FeO:Fe_2O_3$ was not determined; total is therefore reported as FeO. Analyst: F. T. Seelye.

B. Recalculation of analysis A to agree with spinel formula after exclusion of ilmenite and cassiterite.

C. Aluminian chromite separated from mesh-serpentinite, Cerberus Peak, Livingstone Range. Analyst: F. T. Seelye.

D. Recalculation of analysis C to agree with spinel formula; SiO_2, CaO, and TiO_2 neglected.

E. Aluminian chromite separated from mesh-serpentinite, Hapuka Point, D'Urville Island. Analyst: F. T. Seelye.

F. Recalculation of analysis E to agree with spinel formula; TiO_2, SiO_2, CaO and NiO neglected.

G. Chromite from Silver Lease Mine, Oregon, analysis no. 25; 1.3 per cent H_2O + is included in the total. Analyst; R. E. Stevens (Stevens, 1944, p. 11).

H. "Beresovskite"[19], Jimblebah, Western Australia. Analyst: H. Bowley (Bowley, 1940, pp. 203).

Assuming that the spinel structure has a unit cell of 8 ($RO \cdot R_2O_3$) (Bragg, 1937, p. 98), analyses B, D, and F have been recalculated to give the figures in Table 2.

[18] ± 0.003.

[19] Simpson (1920, p. 105) called this chromian spinel, beresofite, a name reserved earlier for chromate of lead. Bowley (1940) has added to the confusion by renaming the spinel "beresovskite", which is quite unnecessary and should be abandoned. Similarly it might be noted that Barsanov (1941) needlessly called chromian magnetite, ishkulite, and listed an analysis in which the $RO:R_2O_3$ ratio is so low that it does not even approximately satisfy the spinel formula. This indiscriminate naming of minerals is to be condemned.

TABLE 2.—STRUCTURAL FORMULAE OF NEW ZEALAND ALUMINIAN CHROMITES (B, D, and F, Table I)

	Wt. per cent	No. of metal atoms	
B			
Cr_2O_3	38.29	7.27	16
Al_2O_3	26.03	7.38	
Fe_2O_3	7.45	1.35	
FeO	13.36	2.68	8
MgO	14.87	5.32	
D			
Cr_2O_3	43.67	8.29	16
Al_2O_3	24.72	6.99	
Fe_2O_3	4.07	0.72	
FeO	11.36	2.26	8
MgO	15.93	5.69	
MnO	0.25	0.05	
F			
Cr_2O_3	43.45	8.26	16
Al_2O_3	24.01	6.79	
Fe_2O_3	5.26	0.95	
FeO	10.90	2.18	8
MgO	16.08	5.79	
MnO	0.18	0.02	
CoO	0.12	0.01	

plotted on this figure, they fall just within the field of aluminian chromite and very close to the boundary of chromian spinel.

The figures for alumina and magnesia appear surprisingly high at first sight, but nevertheless they accord with analyses listed by Stevens and support the view that true chromite (FeO Cr_2O_3) or ferrochromite is of very limited occurrence (Fisher, 1929). On the other hand, it is clear that very extensive ionic substitution is usual in these minerals. To determine whether the composition of the aluminian chromite in the concentrates is characteristic of that in New Zealand peridotites and serpentinites, two samples of pure chromite were separated from ultrabasic rocks from widely separated localities. These analyses (Table 1, columns C and E) are closely comparable with analysis A. So far as the writer is aware, only one other complete analysis of a New Zealand chromite has been made, with a mineral separated from dunite (Peterson, 1869, p. 138); this analysis shows a higher percentage of Cr_2O_3 (56.54 per cent) than any in Table 1 but is too inaccurate to consider further since total iron is quoted as FeO; also the RO group is in excess of R_2O_3 which suggests that some of the iron is in a trivalent state. If the analysis is recalculated to allow for this, the already excessive summation of 101.22 would be increased.

A high percentage of alumina appears to be general in chromian spinels from ultrabasic rocks; Tatarinov and Krasnovsky (1940) have found that alumina averages about 20 per cent in over 60 chromites from harzburgites from localities in the Urals. Further, van der Walt (1942) has drawn attention to similar characteristics for 50 examples of chromites from the Western Bushveld Complex, and Stevens (1944) illustrates similar features for chromites from the North and South American Continents.

Since Bowen (1928, p. 279–281) put forward his views on the origin of chromite and the possibility of considerable variability of composition, little work has been carried out to follow up these ideas. Recently, however, van der Walt's excellent study has shown that the composition of the spinel reflects the composition of the magma from which it separated, and Meniaylov's (1946, p. 352) statement that "chromite has separated out before chrom-spinellid and magnetite" from a complex of basic and ultrabasic rocks certainly suggests control of composition by differentiation. Thayer (1946) has shown very clearly that the composition of chromites and their host rocks may be correlated, and the aluminous nature of the New Zealand aluminian spinels (analyses C and E, Table 1) obtained from mesh-serpentinites of the dunite type with associated plagioclase-bearing ultrabasic and basic rocks supports this.

CINNABAR: An exceedingly rare constituent of the finest screening of one sample only (Barrytown), cinnabar occurs in tablet-like fragments with a perfect cleavage (Pl. 2, fig. 34). The very bright-red color is distinctive but no dichroism was noticeable. When heated alone the mineral volatilized but when heated with sodium carbonate, globules of mercury were obtained, confirming the initial optical diagnosis.

CLINOZOISITE: Iron-poor or iron-free members of the epidote group occur in the majority of the samples studied, limited, with few exceptions, to the finer-grade sizes. Relative frequency counts clearly show that members of this mineral group are of minor importance in the different assemblages of dredge concentrates since they do not exceed 8 per cent of any sieve fraction. In most instances, clinozoisite occurs in subhedral, rather equidimensional grains, although short, stout, prismatic euhedra are well represented in occasional examples (Lowburn). The mineral is colorless with a low birefringence and blue-grey or slate-grey anomalous interference tints; for two crystals the following optical properties were recorded:

$$\beta = 1.728;\ 2V = 87°;\ \rho < \nu$$
$$\beta = 1.723;\ 2V = 83°;\ \rho < \nu$$

According to Winchell's tables (1933, p. 313), these properties indicate approximately 3 per cent of Fe_2O_3 in the crystals concerned. Generally clinozoisite grains are free from inclusions but in some examples, particularly those from Lowburn dredge concentrates, dense clouds of iron-ore dust usually render the grains semi-opaque (Hutton, 1940a, p. 36); this feature of the clinozoisitic members of the epidote group is in marked contrast to the normally inclusion-free condition of the iron-rich members.

CORUNDUM: In colorless or parti-colored anhedra, corundum was a rare constituent of three concentrates (Barrytown, Blackball, and

Gillespie's Beach); quite frequently these grains appear to orient themselves in the mounting media so as to give a centered interference figure, which suggests the occurrence of plates, the shapes of which are determined mainly by the basal parting. A biaxial, negative interference figure with an optic axial angle of 10–20° was not unusual, and a determination of the refractive index for the ordinary ray gave 1.769 ±0.002 for one grain.

EPIDOTE: Epidote has the same general distribution as clinozoisite but in individual samples is a much more important member of the mineral assemblage; in Lowburn and Atarau concentrates it rises to 27 and 10 per cent respectively of the finest screenings (−230 grade-size), and 34 and 23 per cent respectively of the −120 +230 grade-size. In coarser fractions, however, the epidote group makes up only a few per cent of any particular screening, except in the case of samples from Lowburn dredge. Epidote occurs either as highly irregular grains with a general aspect suggesting composite particles, made up of sub-rounded individuals, or as broken fragments showing some tendency to prismatic habit. Typically the mineral is greenish-yellow with a weak but distinct pleochroism and high birefringence giving characteristic high-order interference colors. Inclusions are less frequent than in clinozoisite, but not rare; iron-ore in dust-like state of subdivision or as large particles, grains of quartz and albite, flakes of chlorite and needles of rutile have been identified. In a few grains (Barrytown), zonary structures were found, in which clinozoisite or iron-poor epidote formed cores with the peripheral zones richer in iron.

GADOLINITE: Occasional well-rounded grains of the beryllium yttrium silicate, gadolinite, were recognized in a coarse fraction (−35 +60 mesh) of concentrates from Barrytown dredge[21]. The anhedra are shining-black with vitreous luster and exhibit pronounced conchoidal fracture. Hardness is approximately 6–7 and density 4.25 ±0.03. In thin fragments, the mineral is deep bottle-green, and numerous, minute, opaque particles, not diagnosed, have been seen.

[21] There were fewer than 25 grains in the entire mineral assemblage in the −35 +60 fraction of Barrytown material, and therefore, the individual constituents are not listed in Table 19.

The mineral is quite isotropic, that is completely altered to the metamict state, and has a refractive index of 1.783 ±0.002, comparable to the determination of 1.781 ±0.002 made by the writer for metamict gadolinite from Broddbo, Norway, and Ytterby, Sweden but rather higher than the figure of 1.776 (approximately) found by Mountain (1931) for metamict gadolinite from Onseepkans and Boksput, Cape Province. Incidently the density of the South African material was also distinctly lower than that of the Barrytown example, 4.085, suggesting a greater degree of hydration in the South African mineral. When fragments of gadolinite were brought to dull-red heat, the metamict condition was destroyed and a state of anisotropy resulted. At the same time the color changed to pale brown, and the refractive index rose to approximately 1.80 for the fast direction in the crystal, and a birefringence of 0.015 was measured. However, this transformation did not appear to be accompanied by any sudden emission of light (Levy, 1915, p. 35–36; Faessler, 1942). Hofmann and Zerban (Levy, 1915) have stated that this glowing will take place suddenly at 430°C., whereas Faessler observed that this change occurred suddenly at 800°C. but more slowly at lower temperatures, *viz.* 590°C. Heating of the Westland gadolinite was carried out over a range of temperatures but although the metamict condition was destroyed, a sudden luminescence or glowing was not observed at any stage.

Although ignition to fairly high temperatures will cause metamict gadolinite to return to a condition of anisotropy, it does not necessarily restore the refractive indices to the values that would have been found for the unaltered mineral since the divalent iron is almost wholly oxidized by the heating process, causing a distinct change in color and an increase in the birefringence above that found in the original unoxidized mineral.

In order to verify the determination of gadolinite, Mr. Seelye was asked to check for the presence of beryllium and rare earths in a few grains submitted to him; he supported the writer's identification and reported that "beryllium was identified in the sample by means of the Kolthoff Curcumin reaction after separating rare earths as fluorides, and any excess of

aluminum and iron by means of sodium fluoride. The presence of beryllium was further confirmed by the quinalizarin test."

GARNET: Recognized in every concentrate examined to date and often the predominating constituent in coarse grade-sizes, garnet is rarely of much importance in the finest screenings (Rimu dredge is an exception). The form of the mineral ranges from sharply euhedral crystals to sub-rounded or worn grains, characteristic of the coarser grades, to the sharply angular fragments with marked conchoidal fractures most frequent in the finest screenings. The morphology is always very simple. The rhombic dodecahedron (Pl. 2, figs. 31, 32) is by far the most usual form developed, whereas the trapezohedron is distinctly uncommon; combinations of these forms occur but the former is almost always predominant. In a few examples (Slab Hut), well marked striations oriented parallel to the shorter diagonals suggests oscillatory combination of the dodecahedron and the tetrahexahedron. Although no statistical counts were made, the outstanding predominance of the dodecahedron over the trapezohedral form and the correspondence of composition to almandine with ranging amounts of spessartite in solid solution in all cases except one (a nearly pure spessartite) shows an anomaly in the Donnay-Harker (1937, p. 460–461) law for the crystal morphology of garnet that the trapezohedron should predominate over any other form since the planes bounding this form are parallel to the net planes with the greatest reticular densities (or the smallest mesh areas). However, Pabst (1943, p. 234) has recorded evidence that among 90 almandine garnets studied, the dodecahedron (the form ranking second in importance, according to Donnay and Harker) has slight predominance over the trapezohedron (leucitohedron). For the nearly pure spessartite only the rhombic dodecahedron was observed, in accordance with the writer's observations for similar garnets in the Central and Western Otago schists. Further, a cursory examination of the almandine-rich gneisses from the Southland Fiord country only rarely found any form other than the simple dodecahedron. Thus the marked predominance of the dodecahedron in almandine and spessartites in a number of New Zealand localities seems clear.

In some of the concentrates are occasional garnet fragments of extremely irregular outline (Pl. 4, figs. 57–58) which suggests that they may have been subjected to the action of solvents, but the forms are quite distinct from the authigenic etch patterns found by Bramlette (1929) on garnets from Venezuela, or the pitting and etching observed by Wilgus (1933, p. 86) in garnets from the Dresbach sandstone of western Wisconsin. That garnet may be readily dissolved by natural solvents under some conditions is not universally accepted. Smithson (1942, p. 33–4) believes that garnet in sediments may and does dissolve by means of solvents working along the cleavage planes of that mineral; Edelman and Douglas (1934) concur in this, but believe that the process is rare and that the absence of garnet is more likely to mean absence from the sediments when they were being deposited. L. and C. Dryden (1946) list garnet as even less stable in sediments than hornblende, but Condit (1912) expressed the opinion that garnet was one of the most stable constituents of the mineral assemblages in sedimentary rocks. The extreme angularity and deep re-entrants so well developed in many of the garnets from Snowy River and Rimu could scarcely have survived even the slightest abrasion or transportation; the writer therefore believes that these grains are merely remnants produced by actual solution of much larger grains in the loose, water-filled gravels. Solution with ultimate disappearance of garnet from loose, saturated sediments may possibly be more important than generally considered; clearly this condition need not apply only to older sediments.

A distinct though not wide range of colors was noted in garnets from most of the localities studied, from pale pink, through red to brownish-red; since this range of color would likely be connected with a range of chemical composition, refractive index and density, determinations were made for more than 30 samples from widely separated localities (Table 3). They show a range of 1.795–1.825 for refractive index and 3.94–4.20 for density, Lowburn garnet excepted in both. Thus the pale-pink to pink tints appear to generally characterize garnets with refractive indices usually less than 1.806 and densities that do not exceed 4.01. The deeper red or brownish-red garnets usually have dis-

TABLE 3.—PHYSICAL PROPERTIES OF GARNETS

Specimen	Color	R. I.	Density
Grey River dredge	red	1.825	4.20
" "	pink	1.795	4.01
" "	pink	1.798	3.95
" "	red	1.800	4.08
" "	pink	1.795	3.98
" "	red	1.795	4.02
Snowy River dredge	brownish-red	1.815	4.12
" "	brownish-red	1.815	4.10
" "	brownish-red	1.813	4.10
" "	pink	1.803	3.98
Slab Hut dredge	red	1.809	4.04
" "	red	1.805	4.10
" "	brownish-red	1.805	4.09
Beachsand, 1 ml. from South end, Bruce Bay	pink	1.806	3.98
" "	pink	1.805	3.94
" "	red	1.798	4.00
Concentrate from stream, 1 ml. N. of Lake Paringa.	red	1.795	4.12
" "	pink	1.803	3.98
" "	brownish-red	1.800	4.12
Kaniere dredge	brownish-red	1.820	4.19
Lowburn dredge	pink	1.783	3.90

TABLE 4.—ANALYSIS OF ALMANDINE FROM SLAB HUT CONCENTRATES

SiO_2	36.89	Mode:	
Al_2O_3	20.92	Gr.	2.7
FeO	29.82	Sp.	21.75
Fe_2O_3	nil	Py.	6.0
MnO	9.26	Al.	69.55
MgO	1.82	$RO:Al_2O_3:SiO_2$ =	
CaO	1.06	2.97:1.00:2.99	
TiO_2	0.06		
Na_2O	0.05		
K_2O	0.04		
H_2O at 105°C.	0.15		
	100.07		

tinctly higher densities and refractive indices. Qualitative chemical tests on 10 of the garnets listed in Table 3, showed plentiful manganese. Therefore, these garnets would seem to be chiefly spessartite-almandine with small percentages of pyrope, and less grossularite in solid solution (Winchell, 1933, p. 176). As the garnet in the Lowburn concentrates probably contains in excess of 50 per cent of the spessartite molecule (Hutton, 1940a, p. 24), and is derived from the Chlorite Zone Otago schists,

TABLE 5.—RECALCULATION OF ANALYSIS OF GARNET ON THE BASIS OF 12 OXYGEN ATOMS TO THE UNIT CELL

	Wt. per cent	No. of metal atoms	Arrangement in groups	Ideal composition
SiO_2	36.89	3.00	3.00	3
Al_2O_3	20.92	2.00	2.00	2
FeO	29.82	2.03	2.98	3
MgO	1.82	0.22		
CaO	1.06	0.09		
MnO	9.26	0.64		

Formula: $[Fe_{2.03}Mg_{0.22}Ca_{0.09}Mn_{0.64}]Al_2Si_3O_{12}$.

it is distinct from the other garnets studied here from Westland and Nelson Provinces. In order to determine the chemical nature of the garnets more accurately, a complete analysis (Table 4) was obtained of garnet separated in the pure state from Slab Hut concentrates by electromagnetic and flotation methods. For this material the constants were:

Observed: n = 1.815–1.818
D. 21°C. = 4.18–4.20
Calculated (Ford, 1915):
n = 1.812
D = 4.17

If recalculated on the basis of 12 oxygen atoms to the unit cell and the small percentages of alkalies and titanium are neglected (Table 5), this analysis nearly completely agrees with the structural formula without the necessity of grouping any aluminum with the four-fold coordinated ions, as Alderman (1935) found necessary for a number of almandines from Botallack. Although Fleischer (1937, p. 752) was not able to disprove Alderman's explanation of anomalies in the $RO:R_2O_3:RO_2$ ratio in some garnets, he believed that this position was much more likely due to the difficulties of determining the percentage of ferrous oxide. While fully aware of this very considerable difficulty that besets the determination of ferrous iron in refractory minerals, it is quite clear that in a first-class analysis

of an almandine showing negligible or no ferric iron, the ferrous determination must have been correct. Yet Pabst (1938, p. 101) published an analysis of an almandine with only 0.24 per cent of Fe_2O_3 but with a $RO:R_2O_3:SiO_2$ ratio of 2.82:1:2.87; if this analysis is recalculated on the basis of 12 atoms of oxygen it is only necessary to transfer the almost negligible quantity of 0.02 atoms of aluminum from six-fold co-ordination to four-fold co-ordination in order to bring the silicon group to the theoretical figure of 3. On the other hand there is still a marked discrepancy between theoretical and actual values that cannot be allowed for by appropriate ionic substitution. Similar conditions apply in the case of Pabst's (1943) Fort Wrangell almandine. In both instances, neither ionic substitution nor incorrect ferrous iron determination can explain the anomalies in the oxide ratios. In an almandine from Botallack (Alderman, 1935, p. 46) that is Fe_2O_3-free, correspondence with the theoretical formula results when the appropriate ionic substitutions are assumed; this, however, is not the case of some almandines with small amounts of Fe_2O_3 quoted by Alderman, and the RO group remains too low in spite of a transfer of Fe^{3+} to this group.

Zoning is not infrequently developed in spessartite-almandine, but is not often noticeable unless the particles are immersed in a medium with a refractive index of about 1.80. Two distinct developments were found The more usual variety exhibited numerous concentric bands, in most cases probably less than 1 or 2 microns wide. No distinction in color was apparent but prominent Becke lines separated from these zones on racking the microscope tube up or down. In the second form, the central portions of the garnet crystals were red or brownish-red, but the intensity of absorption decreased without abrupt change towards the margins where a pale pink color predominated. The refringence of the different zones could be determined in only one case: Central zone = 1.795 ±0.003; median zone = 1.806; peripheral zone = 1.795.

Garnet free from inclusions is uncommon and they are usually arranged haphazardly with only an occasional specimen showing evidence that distribution of the inclusions has been influenced by the zonary development (Atarau Dredge); in some crystals inclusions were more densely aggregated in the central areas. Although the inclusions could not always be identified definitely, epidote, clouds of iron oxides or manganiferous dust, magnetite, titanite, quartz, and feldspar were diagnosed. Very deep staining by iron is common.

GOLD: Since samples from dredges had gone through a process of amalgamation, gold was absent except in concentrates from Ngahere and Gillespie's Beach dredges. In Ngahere material gold particles were completely coated with ferric oxides and had to be treated with warm concentrated HCl for identification. Apparently this protective coating of oxide inhibited complete amalgamation, and as a result, valuable amounts of gold were rejected with the tailings. In the Gillespie's Beach dredge, the concentrates may have been imperfectly handled since the gold therein was quite free from any protective coating that would have prevented amalgamation. Gold was found in a number of beach sands and river pannings, and ranged from flattened grains or "slugs" 9 mm. long down to fine flour that passed a 230-mesh U. S. screen. The finest particles are usually skeletal in form but filiform, granular, and flakelike shapes were often found. The metal was readily diagnosed by its distinctive color beneath a binocular microscope in oblique illumination, the ease with which a flake or particle may be pricked and hence picked up on a needle point, and its reaction towards a globule of mercury.

Particles of gold from Ngahere dredge showed a density range of 17.02–18.07.

HEMATITE: In the main hematite is restricted to the coarser fractions of concentrates from Ngahere, Blackball, and Red Jacks, although it is an occasional constituent of a number of grade sizes of a number of other concentrates. Pseudomorphs after pyrite in the form of pyritohedra, striated cubes, and fragments of these are most often found; rolled grains are not infrequent. Very occasionally, notably from Atarau and Slab Hut dredges, thin plates of hematite were found in which the basal plane was dominant but prisms and pyramids exceedingly poorly developed; the density of such crystals ranged from 5.20–5.24. Hematite is dull brown to shining grey in reflected light,

and is opaque except in the thinnest flakes. A considerable number of grains were tested by removing them on a needle point, crushing them directly on a glass slide in refractive index liquids, and then inspecting the powder obtained therefrom in transmitted plane polarized light. These very thin particles were deep brownish-red but not sensibly dichroic. The refractive index was decidedly in excess of 2.08.

Although the writer was unable to obtain any interference figures or to determine the value of the optic axial angle owing to minuteness of particles, the determination of high density and the knowledge that the refractive index is very much in excess of 2.08 for any orientation appears to rule out any possibility of the mineral being lepidocrocite, whereas the high density and absence of dispersion should remove goethite from the list of possible minerals. Therefore, the recognition of these grains as hematite appears to be reasonably well founded.

HORNBLENDE: Few specimens, such as black sands from the western side of Wainui Inlet, north west Nelson, and concentrates from the Kaniere dredge and Arahura River, were found to contain hornblende in excess of 2 per cent. Generally, the amphiboles occur in elongated cleavage plates with ragged ends (Wainui Inlet), in rolled and rounded or angular grains. A green to blue-green type and a rare red-brown barkevicitic variety were noted. In two examples of the former type the following optical properties were found:

Wainui Inlet, north-west Nelson

X = greenish-yellow.
Y = olive-green.
Z = deep bluish-green; almost opaque in thick fragments.
Absorption: strong with $Z > Y > X$.
α = 1.646.
γ = 1.668.
$\gamma - \alpha = 0.022$.
$Z \wedge c = 20°–22°$.

Arahura Dredge, Westland

Pleochroism: similar to the preceding variety, but the refractive indices differ slightly with the γ value showing a range of 1.670–1.683.

The brown varieties usually show strong pleochroism with a range of intensities, the absorption for the Z vibration direction being intense for barkevicitic types.

The hornblende grains from the Wainui Inlet-Ligar Bay sands are completely unaltered but preliminary alteration to chlorite has occurred in some instances (Okarito and Atarau). Inclusions are infrequent but apatite, epidote, opaque minerals, either in clouds of dust-like particles, rods or grains, quartz and feldspars, titanite, and zircon have been definitely diagnosed.

HYPERSTHENE: Hypersthene (Pl. 4, fig. 63) appears to be limited to the finest screenings of one sample only (Snowy River). It occurs in prismatic grains with somewhat ragged terminations and a pale but distinct pleochroism was evident according to the scheme: X = brown to brownish-pink; Z = pale green. Numerous inclusions take the form of (1) minute dark-brown needles that occur in patches and (2) numerous more or less rectangular brown inclusions similar to those figured by Hutton and Turner (1936, Pl. 23, figs. 8, 9). The patches have not been definitely identified but they may be thin plates of rutile lying within and oriented by the planes of the prismatic cleavages. One determination of the refractive index for the Z-vibration direction in hypersthene gave 1.705, which corresponds to approximately 38 per cent of the molecule $FeSiO_3$ (Henry, 1935, p. 223). In all cases the mineral was moderately paramagnetic and could therefore be concentrated by magnetic fractionation.

KYANITE: Kyanite, in relatively few samples and nowhere a constituent of any importance, occurred as rather regularly shaped grains, tabular parallel to (100) and showing two cleavages intersecting at 90° (Pl. 4, fig. 69). An extinction angle of 30° between Z and the 010 cleavage was obtained on this plate, together with a more or less centered biaxial interference figure. Minute inclusions lying parallel to (001) were present in some cases (Red Jacks dredge).

LEUCOXENE: The few rounded grains of leucoxene recognized were always restricted to the finest grade-sizes. The mineral is translucent, pale yellow to amber, with a white porcelain-like appearance in oblique illumination. The refractive index of this constituent, for the grains investigated, was slightly in excess of 2.07, and the birefringence was fairly high. Experiments with a few isolated grains proved that they were anhydrous, lime-free, and very

high in TiO_2. These data are not comparable to those determined by Edwards (1942, p. 273–274) for leucoxene from some Victorian bauxites that he was able to separate pure in quantities sufficient for analysis, since his material contained water and was isotropic.

It would seem likely, then, particularly in view of the X-ray investigations by Tyler and Marsden (1938), Coil (1933), and others, that much so-called leucoxene seems to be merely one modification of titanium dioxide and not titanite as so often assumed. Further, material described as leucoxene appears to be either highly birefringent or isotropic. The isotropic, hydrated, titanium oxides should perhaps be included in the term doelterite, and the term leucoxene restricted to micro-crystalline, anhydrous, highly birefringent titanium dioxide, whether in the form of anatase or not.

MONAZITE: The cerium phosphate, monazite, occurs in all samples examined except those from Central Otago and the beaches of Wainui Inlet. The mineral characteristically occurs in slightly abraded, egg-shaped grains, and in most instances it is concentrated in the −60 +120 and −120 +230 grade-sizes, although often a minor constituent of coarser screenings. Broken fragments are rare. These rounded grains do not give easily recognizable interference figures and are interpreted to be abraded (100) tablets; the (001) cleavage is rarely seen but in some instances a longitudinal cleavage is developed parallel to the clinopinacoid (Pl. 2, figs. 26–28). In rare cases (Pl. 4, figs. 72–73), rectangular-shaped grains show a nearly centered interference figure and a birefringence of about 0.002; such grains appear to be basal cleavage plates, with rectangular form determined by the pinacoidal cleavages.

Although slightly less susceptible than the associated spessartite-almandine garnet and ilmenite, monazite is decidedly paramagnetic, and this allows ready concentration of the phosphate from samples. For instance with a Frantz Separator set at a slope of 15° and a tilt of 8°, practically all of the monazite is attracted at a current of 0.40 amps. A very faint yellow color in the mineral is noticeable only in the coarser screenings, or in electromagnetically concentrated fractions; the color corresponds approximately with Colonial Buff 21″d (Ridgway, 1912). In grains less than 62 microns in diameter the mineral appears to lack color in the usual mounting media, Canada balsam or clove oil.

Refractive index determinations for a number of samples gave the following averages, with none of the measurements differing by more than 0.003: $\alpha = 1.787 \pm 0.003$; $\beta = 1.789$; $\gamma = 1.839$; $\gamma - \alpha = 0.052$. One sample (Ngahere dredge) gave the following data: $\alpha = 1.791 \pm 0.003$; $\beta = 1.792$; $\gamma = 1.844$; $\gamma - \alpha = 0.053$; optical axial angle determinations gave a range of 12°–14°.

Many of the monazite grains contained inclusions but all could not be diagnosed with certainty. The following minerals were found as inclusions: stumpy euhedra of colorless zircon (Grey River, Atarau), granules and dust-like clouds of opaque ores (Blackball), rutile (Snowy River), and red-brown stains (Slab Hut). However, the majority of crystals were devoid of inclusions, and in this respect they are like those described from the Dartmoor area by Brammall (1928, p. 29–30).

The possibility of confusion in diagnosing monazite when such constituents as xenotime, titanite, bästnasite, and yellow zircon are present (Hutton, 1947a) will be considered more fully when discussing xenotime.

The refractive indices of the New Zealand monazites are as low as any recorded for that mineral but their optical properties compare to those for a thorium-free specimen from Bolivia (Gordon, 1939, p. 7), and for monazite with 0.2 per cent of ThO_2 from Katanga (Thoreau, Breckpot, and Vaes, 1936). However, the New Zealand monazites contain over 5 per cent of ThO_2 and 1.23 per cent of U_3O_8 in one instance. Analyses of New Zealand monazites, compared with several others (Table 6), show the presence of 5.48 per cent of the oxides of tantalum and niobium in monazite from Westland. It should be stressed, however, that the sample analysed contained approximately 10 per cent of tantalian-cassiterite, but since this mineral contained only 2.3 per cent of Ta_2O_5, and no other tantalum-bearing mineral was present, the tantalum must be present in the monazite structurally. This is substantiated by the analyst's report that "when the original sample was treated with boiling concentrated H_2SO_4 alone,

the rare earths and phosphorus pentoxide of the decomposed monazite were rendered almost completely soluble and could be removed by filtration, the filtrate containing only a very small amount of the total tin of the sample".

vaal, and Madagascar, but the Ta_2O_5 recorded therein does not exceed 1.10 per cent. If tantalum is, as appears, a replacement ion in monazite its location in the structure is now a question.

TABLE 6.—ANALYSES OF MONAZITE

	A	B	C	D	E	F	G	H	I
Ce_2O_3	19.00	22.95	27.30	28.43	30.72	33.06	25.82	22.63	21.08
La_2O_3 etc.	23.42	28.28	30.00	31.17	30.02	30.21	30.62	34.63	31.27
Y_2O_3 etc.	3.54	4.28	2.24	2.32	2.74	0.14	2.03	4.66	3.53
ThO_2	4.40	5.32	5.28	5.47	5.00	5.03	9.60	7.32	11.08
U_3O_8	1.02	1.23	nil	—	—	—	—	0.32	0.42 (UO_3)
$(Ta,Nb)_2O_5$	4.54	5.48	—	—	—	—	—	—	—
SnO_2 (insol.)	9.50	—	—	—	—	—	—	—	—
SiO_2	2.04	2.46	1.72	1.18	1.20	1.22	1.85	1.54	2.98
TiO_2	1.76	—	0.40	—	—	—	—	—	—
Fe_2O_3	4.62	2.88	1.26	0.93	3.00	2.21	1.01	0.08	0.66
P_2O_5	22.46	27.12	27.50	28.57	26.29	26.70	27.07	27.89	27.52
$ZrSiO_4$ (insol.)	1.08	—	—	—	—	—	—	—	—
H_2O − (105°)	0.22	—	—	—	—	—	0.35	0.06	0.56
H_2O + (105°)	—	—	—	—	—	—	—	0.34	
Undetermined	2.40	—	—	—	—	—	0.18	—	—
CaO	—	—	0.79	0.82	0.15	0.90	0.91	0.35	0.52
ZrO_2	—	—	1.13	—	—	—	—	—	—
Al_2O_3	—	—	1.62	1.03	0.35	0.44	0.15	0.10	0.80
ZnO	—	—	0.50	—	—	—	—	—	—
MgO	—	—	0.08	0.08	—	0.21	0.03	0.02	0.27
MnO	—	—	trace	—	—	—	0.08	nil	—
Loss on ignition	—	—	—	—	0.25	0.59	—	—	—
PbO	—	—	—	—	—	trace	0.58	0.33	—
	100.00	100.00	99.82	100.00	99.72	100.71	100.28	100.27	100.69
Sp. Gr. 20° C.	—	5.26	—	5.23	—	5.26	5.18	5.270	5.17

A. Monazite from Ngahere Dredge concentrate, containing approximately 15 per cent of zircon, tantalian cassiterite, and ilmenite. Analyst: F. T. Seelye.

B. Analysis of pure monazite after elimination of impurities by calculation.

C. Monazite from Mudtown, Pegasus District, Stewart Island, containing approximately 6 per cent of gahnite, zircon, and ilmenite-titanhematite. Total iron ($FeO + Fe_2O_3$) is expressed as Fe_2O_3. Analyst: F. T. Seelye.

D. Analysis of monazite C, after elimination of ilmenite-titanhematite, zircon, and gahnite by calculation.

E. Monazite from Ekole, Northern Nigeria (Johnstone, 1914, p. 57, table 6, analysis I).

F. Monazite from Moolyella (Simpson, 1912).

G. Monazite from Dillingö, Norway (*see* Hintze, 1923, p. 368, analysis IX).

H. Monazite from Dickens Township, Nipssing District, Ontario (Ellsworth, 1932b, p. 26).

I. Monazite from Ishikawa, Iwaki Province, Japan (Shibata and Kimura, 1923a, p. 14–15).

Therefore, 5.48 per cent of $(Ta,Nb)_2O_5$ cannot have been obtained by simple solution of the cassiterite. Tantalum in monazite is not unknown, and Hintze (1923) records analyses of tantalian monazite[22] from Australia, Transvaal, and Madagascar, but the Ta_2O_5 recorded therein does not exceed 1.10 per cent. If tantalum is, as appears, a replacement ion in monazite its location in the structure is now a question.

Since only a small percentage of the tantalum ion occurs, two possibilities present themselves:

(a) Tantalum occurs in four-fold co-ordination replacing phosphorus in the P–O tetrahedra,

or (b) Tantalum replaces ions in the six-co-ordinated groups. This possibility is favored chiefly because:

[22] As distinct from analyses of monazite sands which may include tapiolite, tantalite, tantalian cassiterite, fergusonite, etc.

(1) The Ta–O radius ratio lies within the limits for six-fold coordination although that value is not far removed from the minimum figure when either four- or six-fold co-ordination would be possible.

(2) The writer is not aware of any structures in which Ta^{5+} occurs in four-fold co-ordination.

Tantalum is in six-fold co-ordination in the compounds $NaTaO_3$ and $KTaO_3$ (Rankama, 1944, p. 49) which have the perovskite structure and also in columbite-tantalite (Sturdivant, 1930, p. 88); but in the compound K_2TaF_7, Ta^{5+} is in seven-fold co-ordination (Hoard, 1939, p. 61). However, it must be remembered that any substitution of rare earth ions by tantalum must be very limited in view of the discrepancy of ion sizes.

Substitution of rare earths by quintavalent tantalum and tetravalent thorium involves an upset in electrical neutrality unless there is compensatory exchange elsewhere in the molecule. Significant in this respect is the presence of a moderate percentage of silica in most analyses of pure monazites, although in some instances, such as a specimen from Huron Claim, Manitoba, Canada (Hecht and Kroupa, 1936, p. 98), this oxide may be as high as 8.5 per cent[23]. The monazites from New Zealand show no indication of silica as inclusions of quartz or silicates, or as an alteration product as Ellsworth (1932b, p. 25–26) found. Considerable substitution of phosphorus by silicon can result from the similarity of the radius ratio of silicon and oxygen and of phosphorus and oxygen shown so clearly in the mineral ellestadite (McConnell, 1937). Such ionic replacement will cause an upset in electrostatic balance and, in the writer's opinion, electrical neutrality may be restored, at least in monazite, by substituton of thorium (Th^{4+} 1.10), and tantalum (Ta^{5+} 0.65) for cerium (Ce^{3+} 1.18), yttrium (Y^{3+} 1.06), and lanthanum (La^{3+} 1.22).

The monazite analysed by Hecht and Kroupa contained 8.5 per cent SiO_2 and 14.42 per cent ThO_2, whereas the thorium-free monazite from Bolivia (Gordon, 1939) contained only 0.27 per cent of SiO_2. However, an increase of thorium or of uranium in monazite is not always paralleled by simultaneous increase in the content of silicon (G-I, Table 6). It is not clear how 11.08 per cent of ThO_2, for example, is compensated, certainly it is not by substitution of silicon for phosphorus; this leads one to suspect inclusions of thorium-bearing minerals such as thorianite within the crystals of these monazites.

Table 7.—Recalculation of Analyses of Monazite

	Wt. per cent	Metal atoms	
Analysis B			
Ce_2O_3	22.95	1.294	
La_2O_3	28.28	1.606	
Y_2O_3	4.28	0.350	
ThO_2	5.32	0.186	4.031
U_3O_8	1.23	0.032	
Fe_2O_3	2.88	0.334	
Ta_2O_5	5.48	0.229	
SiO_2	2.46	0.379	3.915
P_2O_5	27.12	3.536	
Analysis B (without Fe_2O_3 and recalculated to 100 per cent)			
Ce_2O_3	23.63	1.333	
La_2O_3	29.13	1.655	
YO_3	4.40	0.360	3.827
ThO_2	5.47	0.192	
U_3O_8	1.27	0.051	
Ta_2O_5	5.64	0.236	
SiO_2	2.53	0.391	4.035
P_2O_5	27.93	3.644	
Analysis D			
Ce_2O_3	28.43	1.616	
La_2O_3	31.17	1.705	
Y_2O_3	2.32	0.191	
ThO_2	5.47	0.193	4.148
Fe_2O_3	0.93	0.108	
CaO	0.82	0.131	
Al_2O_3	1.03	0.186	
MgO	0.08	0.018	
SiO_2	1.18	0.184	3.939
P_2O_5	28.57	3.755	

Analyses of both New Zealand monazites have been recalculated on the basis of 16 oxygens to the unit cell (Parrish, 1939) and set out in Table 7. Silicon has been grouped with phosphorus and it is particularly evident that without including 0.379 and 0.184 atoms of silicon in analyses B and D the four-coordinated groups would be distinctly less than the struc-

[23] The high summation of their analysis, 101.14 appears to result from calculation of cerium and rare earths as CeO_2 etc.

tural value required. On the other hand, the ferric oxide in monazite analysis B, 2.88 per cent, is believed to occur structurally and not result from chemical solution of inclusions of ferriferous minerals. This contention is supported by the second calculation of analysis B from which ferric oxide was excluded and the value for the six-coordinated group is distinctly less than 4. In analysis D, the small percentages of CaO, Al_2O_3, and MgO were included with the main constituents.

OLIVINE: A beach concentrate recently sent the writer from Bruce Bay, South Westland, the analysis of which is not listed in Table 19, differed from other samples from that locality because olivine was found in the −35 +60 size fraction. This mineral occurred in chipped, clear bottle-green unaltered fragments with pronounced conchoidal fracture. The color was most obvious when the particles were studied in oblique illumination with the binocular microscope, but the crushed fragments were colorless in transmitted light. Olivine is completely devoid of iron oxide inclusions but nevertheless was concentrated in the more paramagnetic fractions along with monazite. Refractive indices determined for a number of grains showed the following range: $\alpha = 1.670 - 1.678$; $\gamma = 1.704 - 1.712$.

OPAQUE MINERALS: The opacity of a number of the minerals has rendered diagnosis difficult; it should be noted, however, that two of these minerals, martite and wolframite, are translucent but only in exceedingly thin plates or splinters, such as are produced when individual grains are pulverized.

Several grains belonging to this group, but not satisfactorily identified, owing chiefly to rarity of material, are not discussed. However, a few grains found in the −35 +60 mesh fraction of concentrates from Blackball Dredge and classified as "unidentified" (Table 19) gave strong reactions for Fe, Mn, and Ta; these data in conjunction with relative insolubility in strong acids, marked cleavage, and high density, 7.2–7.4, suggest the possibility of a member of the columbite-tantalite series, close to the tantalum end of the series.

Minerals definitely identified are: ilmenite, magnetite, martite, wolframite, ilmenite-titanhematite intergrowths, chalcopyrite, pyrite, marcasite, lead, arsenopyrite, pyrrhotite. A few grains with properties that render them distinct from these listed could not be definitely identified, but they do not exceed 0.5 per cent in any one sample, or fraction of a sample.

Identification of minerals that are opaque to light even at diameters of approximately 50 microns is no easy matter. It is admitted that such material may be embedded in cement, polished, and then diagnosed in vertical illumination by its reflectivity, reaction to chemical solutions, and so on, and this method has been used in a few instances, but unfortunately from a practical point of view, when an investigator is dealing with hundreds of samples, each of which may be fractionated into twelve or more parts, the labor involved in the work of mounting and polishing is wholly excessive and is unwarranted if the percentage of such material appears to be non-economic. But to identify the opaque constituents with the aid of binocular microscope and oblique illumination alone proved to be most useful[24].

Ilmenite and magnetite are ubiquitous in the paystreaks of beach sands and rivers, and in artificially produced concentrates alike, and in almost every sample ilmenite predominated over magnetite. In a number of dredge concentrates and stream-pannings, however, a higher percentage of magnetite was present than in any of the naturally concentrated beach sands except one, a buried pay-streak at Gillespie's Beach that contained 67 per cent. The higher concentration of magnetite in dredge concentrates is of interest and it should be remembered that dredges' jigs and tables are set as near the optimum positions as possible to allow retention of gold with a density of about 19. Since it is decidedly more dense than ilmenite (density *ca.* 4.79) magnetite (density *ca.* 5.175) might be expected to be retained to a somewhat greater extent. Generally speaking, grading analyses of sands, pannings etc. indicated that the frequency of these two constitutents, and also other opaque minerals, increased with increase of grain-size; the opaque minerals, chiefly ilmenite and magnetite, were the predominant con-

[24] A finely crushed opaque powder, e.g. pyrite, can often be diagnosed more easily in oblique illumination, if it is covered with clove oil which makes the color of the pyrite more obvious. The properties of some minerals when viewed by vertical illumination are more easily recognized when the surface is covered with a thin film of clove oil, especially if that surface is not very well prepared.

stituents in the −35 +60 screening of about half of the dredge concentrates investigated.

The very marked predominance of ilmenite over magnetite is of some importance since it had been believed that with the establishment of a steel industry in New Zealand, the West Coast blacksands could be utilized to augment the Parapara limonite and hematite ores (Mason, 1945). With one exception, a blacksand layer 14 feet below the present level of Gillespie's Beach, not one sand contains magnetite in economic quantities (Fig. 5).

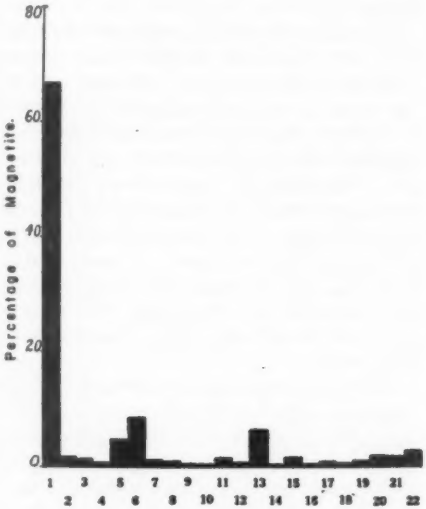


FIGURE 5.—MAGNETITE IN WESTLAND AND SOUTHWEST NELSON BEACH SANDS

1. 14 feet below surface, Gillespie's Beach; 2. 3 miles south of Poerua River; 3. South end of Hunt's Beach; 4. Bruce Bay; 5. Center of Three Mile Beach, Okarito; 6. 200 yards north of Five Mile Bluff; 7. Five Mile Beach, Okarito; 8. Ross; 9. 13 miles north of Greymouth; 10. 14 miles north of Greymouth; 11. 15 miles north of Barrytown; 12. West of Charleston Hotel; 13. Karamea Beach; 14. One mile south, mouth of Buller River; 15. Tauranga Bay; 16. North end of Nine Mile Beach; 17. Nine Mile Beach, south of Westport; 18. 1 mile north of Cape Foulwind; 19. Greysand, 1½ miles north of Cape Foulwind; 20. Mouth of Fox River; 21. Blacksand, 1½ miles north of Cape Foulwind; 22. Dune sand, Bruce Bay.

Both ilmenite and magnetite occur as rolled, well-rounded grains, angular fragments with conchoidal fracture, and as euhedra. In many instances anhedral grains of the two minerals could not be differentiated by visual means alone. One of the better methods of distinguishing between them is to pass the particles through a weak magnetic field. Magnetite is attracted and ilmenite remains behind. If the field strength is then increased slightly opaque grains with a distinct range of magnetic susceptibility can be attracted before pure ilmenite is removed. To determine how far variation in titanium content accounted for this range, analyses were made (Table 8), in some instances partial only, of four fractions; no attempt was made to separate out different opaque minerals but translucent minerals were removed as far as possible, chiefly by hand-picking.

TABLE 8.—ANALYSES OF OPAQUE MINERAL FRACTIONS FROM NGAHERE DREDGE CONCENTRATES

	A	B	C	D
SiO_2	4.05	3.59	0.7	1.4
TiO_2	42.44	20.31	6.8	1.6
FeO	37.18	21.98	29.6	30.0
MnO	1.68	0.5	1.1	0.3
MgO	0.56	—	—	—
Fe_2O_3	8.17	48.15	60.4	63.4
Al_2O_3	3.15	3.47	0.9	1.2
Cr_2O_3	0.45	—	—	—
V_2O_3	0.016	—	—	—
H_2O −	0.54	0.78	0.3	0.3
Zircon & insol.	1.90	0.92	0.4	2.4
	100.13	99.70	100.2	100.6
D. 25°C.	4.61	—	—	—

A. Least susceptible fraction consisting of nearly pure ilmenite; solution in acids difficult. Analyst: F. T. Seelye.

B. Moderately magnetic fraction, consisting of martitized magnetite and ilmenite. Analyst: C. Osborne Hutton.

C. Strongly susceptible fraction composed chiefly of titanomagnetite. Analyst: C. Osborne Hutton.

D. Most susceptible fraction; chromite removed by hand-picking. Mainly magnetite. Analyst: C. Osborne Hutton.

Crystal form has been rarely preserved in beach sands but is more often found in the river concentrates, and served at times to distinguish between the two minerals. Ilmenite usually had squat tabular crystals produced by development of prominent basal plane and rhombohedrons, but ranged to grains with remnants of basal planes and rhombohedrons

and most commonly of all, anhedral grains with hackly fracture. A considerable number of anhedral grains tended to be either prismatic or distinctly platy. Magnetite showed the same wide diversity of grain-shape, but, when euhedral, octahedral crystals were more frequent than octahedrons modified by dodecahedral faces; cubes modified by octahedron faces were noted but rarely.

Since magnetite grains began their sedimentary evolution as octahedra it is perhaps to be expected that abrasion of such crystals would lead to the development of anhedral grains with a greater degree of sphericity than ilmenite. This distinction is quite marked in the river concentrates but is much less so in the beach sands. A conchoidal fracture and a shining metallic luster were much more noticeable in magnetite than in ilmenite (Milner, 1940; Raeburn and Milner, 1927), and although a number of writers have stated that ilmenite has a steel-grey luster with a distinct purple sheen or iridescence, this property was observed just as often in magnetite.

A number of grains of magnetite showed a distinctive set of parallel striations but these could not be diagnosed further. In addition to the purple iridescence, magnetite alone exhibited a red-brown coating of oxide, and in river sands in particular this feature provided quick identification.

While discrimination between ilmenite and magnetite by mere inspection is usually difficult and often impossible, careful observation for degree of luster, well-developed conchoidal fracture, striations, red-brown oxide coating, and form in euhedra is often of considerable assistance. Of course final testing beneath a binocular microscope in oblique illumination with a fine magnetized needle should assist accurate diagnosis. Unfortunately this method is quite inadequate at times; for example octahedra of partially martitized magnetite recognized in a number of sands have in some cases magnetic susceptibility approximately equal to that of ilmenite.

Martite, found in a number of sands and concentrates, reached its greatest abundance in the -35 $+60$ screening of the concentrates from Lowburn Dredge. It ranged from well-rounded or fragmental anhedra to octahedrons and dodecahedrons or combinations of these. In the latter, striations parallel to the junctions of the octahedron-dodecahedron faces were developed. Distorted cube forms and pentagonal dodecahedrons were found, and in the majority of cases the faces of the cubes were strongly striated so as to produce a pyritohedral symmetry.

In oblique illumination, martite was iron-black with a splendent metallic luster in most cases although occasionally dull. Each particle was tested for magnetic susceptibility, removed on a needle point crushed on a glass slide in a liquid of refractive index 2.07, and then inspected in transmitted light with the polarizing microscope. When the refractive index of the crushed powder, which was deep red-brown in transmitted light, was greatly in excess of that of the liquid, it was assumed that the material was hematite and not goethite. The martite appears to be pseudomorphous after magnetite or pyrite.

The pseudomorphs after pyrite must be distinguished from the equally abundant pseudomorphs of similar form that when crushed give yellowish-brown powder composed of cryptocrystalline and amorphous material; these are diagnosed as limonite (used *sensu latu*) pseudomorphous after pyrite.

Roughly prismatic and angular grains of *wolframite* were usually readily recognized beneath the binocular microscope by the presence of perfect cleavage, together with the pronounced and brilliant reflection of light from the cleavage planes. The particles are opaque and, by reflected light, iron-black. If, however, a grain is crushed, the fragments must be washed free from fine dust, otherwise the particles appear to be a deep red-brown in reflected light. In transmitted light, very thin flakes of the mineral were deep red-brown with only faint pleochroism according to the scheme: X = deep red-brown; Z = deeper red-brown to opaque in thicker flakes. The refractive index in every instance was distinctly greater than 2.07. The pleochroism suggests that the wolframite has a composition close to that of ferberite. This was checked by making colorimetric determinations of the manganese content using small fragments of the mineral; the values ranged between 3.2–4.3 per cent of MnO, and according to Hess and Schaller (1914), would indicate ferberite rather than hubnerite. Incidently the compo-

sition of Westland wolframite appears to compare closely with wolframite from Port Pegasus, Stewart Island (Williams, 1934, p. 353), and found to contain 19.6 per cent of FeO and 3.9 per cent of MnO in one instance.

Wolframite was found to be relatively insoluble in hot concentrated hydrochloric acid and repeated evaporations of the finely powdered mineral in boiling aqua regia were needed to produce any sign of the canary-yellow oxide WO_3.

Associated with wolframite in a number of Westland sands and with the wolframite-gahnite-cassiterite-monazite assemblage at Port Pegasus is a mineral that is superficially comparable to wolframite in many ways (Table 9). In reflected light these mineral grains showed what appeared to be good cleavage, are dark brown to black, and very thin flakes have a similar appearance to wolframite. The mineral has approximately the same hardness as wolframite and a dark red-brown streak, but a density of approximately 4.8. The magnetic susceptibility of the mineral was about equal to that of wolframite. Interpretation of the analysis (Table 9) was obscure until a polished section made of a few grains from the $-18 + 35$ screening of Blackball dredge and Mudtown concentrates proved that the material under investigation was of a composite nature and probably an intergrowth of ilmenite and hematite. A recalculation of analysis of the Mudtown material on the assumption that the hematite is titanhematite containing in solid solution the maximum allowable 10 per cent of TiO_3 (Ramdohr, 1926, p. 357), and the ilmenite is saturated at 6 per cent of Fe_2O_3 (Edwards, 1938, p. 42), gives a mode of approximately 46 per cent of titanhematite and 54 per cent of ilmenite with no excess of oxides; the small percentage of the pyrophanite molecule was included in ilmenite rather than with titanhematite. In similar ilmenite-titanhematite intergrowths, Edwards (1938, p. 46) mentions the development of pronounced lamellar twinning; although this feature was not present in the polished section of the New Zealand minerals, the pronounced striations so clear in reflected light may well have been this and not cleavage as supposed.

It is believed that much so-called arizonite (Palmer, 1909; Wadia, 1943, p. 10) and also minerals similar to that described by Mackintosh (1885, p. 342–3) are probably of this nature.

Ragged anhedral grains believed to be *chalcopyrite* have been observed as rare constituents of the coarser fractions of Ngahere dredge concentrates. The mineral was quite opaque even in the thinnest flakes and the finest powder but in oblique illumination has a bright brassy yellow color. Green and greenish-blue staining of this material was very rarely seen, but a brown limonitic coating was more prominent. The mineral is readily soluble in hot concentrated nitric acid.

TABLE 9.—ANALYSIS OF ILMENITE-TITANHEMATITE INTERGROWTH FROM MUDTOWN, STEWART ISLAND
(Analyst: F. T. Seelye)

Fe_2O_3	45.48
FeO	20.28
TiO_2	30.50
Al_2O_3	1.00
MnO	1.49
MgO	0.21
CaO	nt. fd.
SiO_2	0.52
H_2O at 105°C.	0.21
	99.69

Pyrite was found in most samples from the dredges but in few beach sands; and in the latter category an important sample was collected just north of the Waitakere River mouth in the Nelson Provincial District. In most examples the percentage of pyrite in the mineral assemblages of the various screened fractions increased with increase of grain-size, and in one instance, the $-10 + 18$ fraction from Ngahere Dredge, pyrite was predominant.

In the finest grades, pyrite ordinarily occurred in anhedral and splintery fragments and in a few instances (-230 fraction, Snowy River) as globular or concretionary particles; in one sample dendritic forms were common (Blackball dredge). Between these forms and the euhedra so characteristic of the coarser fractions of many samples, there is every gradation.

In the coarser fractions, cubes, perfect, distorted, or striated, pentagonal dodecahedrons (pyritohedrons), and combinations of these were

the most usual forms. Rarely cubes modified by slight development of octahedron faces were found. Pentagonal dodecahedrons with lineage structures (Buerger, 1932, p. 181) were not uncommon.

When the mineral is quite fresh and unaltered, it has a characteristic pale brass-yellow color but in many examples this was obscured by the brown tarnish of oxidation. Such crystals when crushed were frequently found to be completely pseudomorphed by limonitic material or hematite, but however complete the alteration, the detection of parallel striations, which result from oscillatory combination of (100) and (120), was usually sufficient evidence of pyrite, or a pseudomorph thereof. Anhedra of pyrite are sometimes difficult to distinguish from similar particles of chalcopyrite by mere inspection but two tests were usually sufficient: (a) approximation of hardness by crushing, and (b) addition of ammonium hydroxide to the nitric acid solution of the mineral gives a deep blue coloration if the mineral is chalcopyrite; $Fe(OH)_3$ is of course precipitated in both cases.

No *marcasite* was diagnosed with certainty and the problem of distinguishing between it and pyrite when they occur in grains devoid of any crystal faces deserves some comment. Bannister (1932, p. 180) has pointed out that most text-books base description of the color of marcasite and pyrite on tarnished specimens; in freshly polished specimens, marcasite is tin-white and pyrite is pale brass-yellow. In addition, marcasite is strongly anisotropic but pyrite is isotropic or exhibits only very faint polarization colors. Solution of the powdered mineral in hot concentrated nitric acid, suggested by many writers (Ford, 1932, p. 438) is not a reliable method of distinguishing between the two minerals. In doubtful cases, polished sections should be made and the color of the mineral and its effect on polarized light determined. These observations should suffice for accurate diagnosis, but there is a minimum limit to the grain-size that may be satisfactorily tested in this way.

Lead, though absent from the beach sands, is a constituent of a number of dredge concentrates, and it was limited to mineral assemblages coarser than 250 microns. In a sample from Blackball Dredge, lead made up 8 per cent of the $-35 + 60$ fraction. The metal was invariably found as spherical pellets dulled by thin gray-white surface coatings. Freshly cut surfaces had a high reflectivity and the material was completely and readily soluble in warm dilute nitric acid.

There is every reason to believe that the lead in the different samples has been derived initially from shot-gun cartridges, the result of sportmens' activities within the immediate drainage area, and that it is not native lead. The metal never occurred as slugs or aggregates but appeared to be limited to particles within a narrow range of size. Arsenic is present in each of the particles tested, and generally about 0.2 per cent of arsenic is added to lead during the manufacture of shot for hardening purposes. So far as the writer is aware arsenic is not a contaminant of native lead.

The presence of *arsenopyrite* was confirmed in one specimen only, where it was found to occur as rare subhedra with striated prisms and a steel grey color when observed in oblique illumination.

Pyrrhotite is another sulphide that is restricted to dredge concentrates in its distribution, and, when present in an assemblage, only occasional particles were found. The mineral occurs as irregularly shaped anhedra and less commonly as spherical aggregates that seem to be composed of finely divided pyrrhotite cemented with a clay-like matrix. The following properties taken together are considered to be diagnostic: (1) Reddish-yellow to bronze color in oblique illumination; (2) Strongly ferromagnetic; (3) Soluble in warm hydrochloric acid with evolution of hydrogen sulphide; (4) Some grains were determined by means of the dimethylglyoxime test to be nickeliferous.

ORTHITE: Although the cerium epidote is not an uncommon constituent of the granites and schists in the areas from which the sands studied were originally derived, orthite was a very rare constituent of those sands and appears to be limited to Snowy River, Gillespie's Beach, Arahura, and Wainui Inlet material. Orthite occurred as slightly chipped, prismatic crystals, transparent, deep brown and distinctly pleochroic; the mineral has a birefringence of about 0.020 with $\beta = 1.730 \pm 0.003$, for Wainui Inlet material.

The author is of the opinion that some of the semi-opaque, brown-colored, anhedral grains

found in a number of preparations and not satisfactorily diagnosed might well be orthite. Since that mineral is often found in various stages of metamictization and a number of other radioactive minerals have similar physical properties, recognition is often difficult, if not impossible, unless sufficient material can be separated for chemical tests.

RUTILE: Rutile has been observed in most of the finer fractions of dredge concentrates but is generally absent from beach sands; on occasions the mineral has been found in screened fractions coarser than 35 mesh, *viz.* Slab Hut. Stumpy, prismatic, subhedral grains with rounded extremities are usual (Pl. 4, fig. 75), with broken euhedra and well abraded spherical grains not uncommon. In those grains showing a minimum of abrasion, longitudinal striations parallel to the prism edge were often seen, although occasionally striations oblique to the elongation were observed. Geniculated fragments are rare.

Color is typically reddish-brown with a poor dichroism as follows: O = yellow to golden-yellow; E = red-brown; E > O. Rutile from Barrytown dredge was similar except that the ordinary ray has a distinct greenish-yellow tint. A foxy-red color for this mineral was very rarely observed.

Some rutile crystals of one sample (Slab Hut) were separated for determination of density and values ranging from 4.24–4.26 were found with the Berman balance. These data indicate that such rutile crystals, which incidently are comparable with those in other samples, are pure TiO_2, or that the ions that may substitute for Ti, *viz.* Fe, Ta, and Cb, are present in traces only.

In oblique illumination, rutile is readily distinguished from the associated minerals by its prismatic habit, brown color, and resinous luster; however, it is difficult to discriminate thus between broken or anhedral grains of rutile with pronounced sub-conchoidal fracture, and lighter-colored cassiterite fragments. Distinction may be made between such grains by removing them on a needle point, crushing them in oils, and examining the powder by transmitted light.

SCHEELITE: Scheelite is a characteristic detrital mineral in many of the concentrates from Southern Westland and particularly noteworthy in samples from: Five Mile Beach, Gillespie's Beach, Harihari, Saltwater Creek, and Bruce Bay. From other localities in northern Westland, southern Nelson etc., scheelite was either rare or absent altogether. In the southern Westland occurrences, scheelite is most abundant in the $-60 + 120$ screenings but is almost restricted to these and finer material.

Scheelite occurs most frequently as equidimensional anhedra, which in oblique illumination often exhibited bright reflections from octahedral and possibly basal cleavage faces. Occasionally, however, extraordinarily beautiful sets of pyramidal faces were developed (Pl. 2, fig. 37), as many as 45 complete pyramids having been counted on a grain measuring approximately 90 microns in diameter. From the position of the cleavage faces, the pyramidal faces were interpreted as p (111) forms, but no tripyramidal forms were obvious. In those crystals in which pyramidal forms were well developed, the apices of the crystals were almost invariably truncated by a small plane (Pl. 2, fig. 37); it is not clear whether this is an expression of basal cleavage or represents a poor development of c (001), but the latter possibility is favored.

The removal of the apices of the pyramidal crystals might naturally be ascribed to damage during transportation but for the regularity and evenness of the development. Between these and the anhedral fragments are grains (Pl. 2, figs. 35–36) showing every gradation in the perfection of form.

Since the optical properties of scheelite are so similar to those of a uranium-free thorium silicate, tentatively called thorite, found associated with the tungstate, and since preliminary examinations failed to differentiate between them, the physical properties of both minerals have been determined with care. To enable this to be accomplished satisfactorily about 2 grams of scheelite-thorite concentrates were prepared from original material by screening and removal of magnetically susceptible material[35]; after splitting, convenient quantities were mounted in Canada balsam and hyrax. The following properties were determined for scheelite: Uni-

[35] The southern Westland uranium-free thorite may be removed from a mixture with scheelite but only in an intense magnetic field; with the Frantz Isodynamic separator, fractionation was made at 1.4 amps., cross slope 10°, and longitudinal slope 15°.

axial and optically positive; O = 1.919 ± 0.001; E = 1.935; E − O = 0.016; no dispersion; D.20°C = 6.03± 0.02.

The mineral is very pale yellow-white and, in oblique illumination, a sub-vitreous luster is characteristic. Apart from its optical properties,

TABLE 10.—ANALYSES OF RUTHENIAN SISERSKITE

	A	B
Os	49	41.9
Ir	35	34.8
Ru	10	11.8
Rh	2.5	0.1
Pt	trace	8.1
Fe	—	2.3
Undetermined	3.5	—
	100.0	99.0
D. 23°C	19.6	—

A. Kaniere River Dredge, Westland, New Zealand.

B. Nevyansk, Ural Mountains, Russia. Analysis no. 4, table 3, Swjaginzeff, 1932, p. 182.

scheelite may be readily distinguished from the optically similar thorite by the yellow WO_3 that is formed on its grains when boiled in aqua regia for 3–4 minutes; further, in ultraviolet light of wave-length approximately 2540 Å., scheelite gives a characteristic blue fluorescence whereas thorite emits a faint dull white with a pink tinge.

SILLIMANITE: Sillimanite has been recognized in the concentrates from Snowy River and Rimu Dredges. In both instances the mineral was found to be restricted to screenings coarser than 230- and finer than 120-mesh and only occasional grains were found in either sample.

In both occurrences, sillimanite took the form of short, stout, colorless prisms with pronounced cross-fracture (Pl. 4, figs. 70–71); the crystals are devoid of inclusions, have parallel extinction and positive elongation. Determination of optic axial angle in one crystal by universal stage methods gave a value of 25° ± 3°.

SISERSKITE: A ruthenian variety of siserskite was not identified in any of the samples examined by the writer but it is listed here, since several large oval plates or slugs weighing 2.7 gms. in all were recovered on one occasion during a wash-up of the final concentrates from Kaniere Dredge by the dredge-master. One of these plates has been analysed by Mr. F. T. Seelye (Table 10, analysis A).

The analyst further reports: "Palladium was not detected and platinum if present is probably in very small amount." In a thorough search of the literature available to the writer an analysis of siserskite from the Ural Region listed by Swjaginzeff (1932, p. 182) was the closest approximation found (Table 10, analysis B). In most analyses of iridosmine discussed by Swjaginzeff, platinum is present, with one analysis showing as much as 13 per cent of that metal. Platinum, however, is present in all analyses of siserskite, except one (1932, p. 174, analysis no. 9), an old analysis which may not be quite trustworthy. The lack of platinum in the New Zealand material is, therefore, somewhat unusual.

The metal has a dull silvery-grey luster and steel-grey to white color in oblique lighting. Even in naturally rolled and worn grains, a distinct cleavage is evident. It is relatively hard and cannot be pricked and picked up with a needle point. This property, together with siserskite's insolubility in aqua regia and distinct cleavage, constitute useful criteria for distinguishing this alloy from platinum.

The name siserskite is used here in its original sense, since in the New Zealand material Os > Ir (Palache *et al.*, 1944, p. 113).

SPINEL: A ferroan variety of spinel is an occasional constituent of the finest grade-size of material from Red Jacks dredge concentrates, occurring as dull-green fragments with pronounced conchoidal fracture that are translucent when finely crushed. The mineral is isotropic, devoid of inclusions but intersected by numerous cracks; it is moderately paramagnetic and has a refractive index of 1.782 ± 0.002 and a density determination gave 3.74 ± 0.01.

The writer would like to stress the fact that before material from this dredge was studied, a few fragments of gadolinite had been identified from Barrytown material, and when the granules of spinel were found, it was by no means obvious at first that it was green spinel, due to the optical similarity of gadolinite and ferroan spinel. However, the densities of the two minerals are sufficient to allow differentiation,

and more noteworthy still, upon heating, spinel will remain isotropic whereas gadolinite becomes strongly anisotropic since its metamict condition is destroyed. Further confirmation was obtained by microchemical tests for the presence of magnesium. These details are stressed because it has been observed that, in California, students would often quickly dismiss an isotropic green detrital mineral as chrome garnet simply because that mineral has been diagnosed in the heavy mineral assemblages of sediments from that State, when instead they were dealing with green spinels. Thus green isotropic grains should be identified with some care, but usually refractive index data are sufficient to lead to correct identification. Minerals such as yttrialite and gadolinite, which are usually found in a metamict state and have properties similar to those of green spinels, cannot be recognized by inspection alone.

STAUROLITE: Only rare anhedral grains of staurolite were found in the −230 fraction of the concentrates from Blackball Dredge (Pl. 4, fig. 74). Since this mineral has not previously been recorded from New Zealand, the grains were isolated and all the data possible were determined for them, as follows:

α = 1.744 ± 0.003
γ = 1.755
$\gamma - \alpha$ = 0.011
Biaxial and optically positive.
2V = large, probably in excess of 75°.
Pleochroism is weak according to the scheme:
X = colorless to very pale yellow.
Z = pale yellow.
Z > X

The mineral exhibited no cleavage, but had a pronounced conchoidal fracture and numerous sinuous lines of minute opaque inclusions that are believed to be carbonaceous material.

THORITE: Several varieties of thorite have been noted in specimens from a wide variety of localities, but they were never more than mineralogical curiosities except in the central and southern Westland region. In concentrates from Harihari, Saltwater Creek, Gillespie's Beach, Okarito, and Bruce Bay areas, a biaxial thorium silicate, hereafter called thorite, and uranothorite sometimes ranked as the most important minerals present after removal of iron ores. Since three distinct varieties of thorite have been found each will be discussed separately.

Uranothorite is particularly abundant in the −120 + 230 mesh fractions of samples from Harihari, and this material is more or less typical of all other occurrences. It (Pl. 3, figs. 46–54) occurs in well worn and rounded grains to strictly euhedral crystals of simple habit in which the only forms developed are *m* (110), and *p* (111); occasionally poorly developed basal planes were observed (Goldschmidt, 1923, vol. 8, table 80, fig. 4). Splinters and fragments with marked conchoidal fractures were not uncommon. Generally the mineral grains were quite translucent but in a number of cases they were rendered semi-opaque by dense aggregates and liberal dustings of unidentifiable alteration products. In many instances, crystals of uranothorite exhibited curiously anastomosing systems of cracks with no orderly arrangement, although occasionally they appeared to originate in the center of a crystal and to radiate outwards. In some crystals, the central areas were somewhat darker than the remaining portions, and still less frequently, contained dark or opaque inclusions. Not uncommonly the walls of these cracks were stained with brown films, possibly limonitic, and in rare instances these spaces were filled with opaque dust-like particles determined as magnetite. Whether the limonitic stains are the result of alteration of uranothorite is not clear, but the strings of magnetite dust would certainly seem to have been so derived. These irregular networks of cracks have often been observed in radioactive minerals (Brögger, 1890, p. 123–126, in thorite and orangite; Ellsworth, 1927a, p. 367, in euxenite; Hess and Wells, 1930, p. 19–20, in samarskite.); and, further, in many instances have been found to radiate outwards from the actual mineral grains into the surrounding matrix of crystals[36].

There is considerable variation in color and it is difficult to match them exactly with the

[36] Hidden, 1905, p. 423 (surrounding cyrtolite); Landes, 1932, p. 386 (surrounding allanite and fergusonite); Spence, 1930, p. 480; Richmond, 1937, p. 293 (surrounding orangite and allanite); Walker and Parsons, 1923a, p. 17 (surrounding ellsworthite, zircon, fluocerite, and allanite); Ellsworth, 1924, p. 12cI–14cI (surrounding zircon, cyrtolite, allanite, hatchettolite, columbite); Henrich, 1948, p. 68 (surrounding euxenite); Shannon, 1926, p. 36–37 (surrounding hatchettolite).

Ridgway standards (Ridgway, 1912). However, they range from drab, dirty brown (c.f. Ridgway's Isabella Color, 19″ i) to olive-yellow (c.f. Ridgway's 23″ olive yellow). In a few crystals the color is closer to Ridgway's deep chrysolite green (27″), and these are usually quite translucent and free from dusty alteration products. By and large, the more translucent the crystals the more decidedly green their color.

Uranothorite is distinctly dichroic, but the primary colors depend upon the degree of alteration of the mineral. The clear, translucent crystals show the most pronounced dichroism as follows: X = pale lime green (c.f., lime green 25″); Z = Courge green, 25″, i.; Z > X.

For the analysed material the following range of refractive indices was measured: α = 1.818–1.825 ± 0.003. γ = 1.839–1.840, γ-α = 0.015–0.021. For separate grains the lowest value found for α was 1.815, whereas the highest figure for γ was 1.850. For a considerable number of crystals the actual determination of α and γ was impossible, or rendered exceedingly difficult, owing to an apparent recrystallization of uranothorite to give radiate aggregates of flake-like individuals, though still retaining the simple tetragonal prismatic habit and outline. These aggregates, oriented in all possible directions, give such crystals anomalous optical character since, during a 360° rotation of the microscope stage, total extinction was never completely obtained. For clear green homogeneous uranothorite, uniaxial and optically positive interference figures were obtained.

Owing to the high density of uranothorite, flotation methods were impractical for determination of this property. Measurements were made with the Berman balance, but since the grain diameters did not exceed about 0.2 mm., it was not possible to employ single grains or to differentiate between altered and fresh material. Using approximately 8 mgm. portions of pure uranothorite, several determinations gave an average value of 6.7 ± 0.05. A similar figure was obtained using a micropyknometer.

The mineral was moderately paramagnetic but with a susceptibility decidedly less than that of monazite, and as a result uranothorite was fairly easily fractionated from its associated minerals[27]. A nearly pure sample of uranothorite prepared by the writer and analysed by Mr. F. T. Seelye (Table 11, analysis A) shows that the mineral is a true uranium-bearing thorite, which is decidedly low in water-content in comparison with uranothorites from other localities. This particular feature is reflected in the high density and anisotropic state of the Westland mineral and is in marked contrast to the thorite from MacDonald Mine (analysis C), which has a much lower density and is completely metamict. The dominant green of Westland uranothorite is also distinctive since most thorites, whether they contain uranium or not, are red or brown.

A *uranium-free thorite*[*] has a similar general distribution to that of the uranothorite but it is somewhat more abundant. In marked contrast to the associated uranothorite, thorite is colorless to palest yellow; appears to be almost completely non-magnetic, contains no uranium and is almost always anhedral; an approach to subhedralism was rarely observed in any preparation and the grains generally were highly irregular in outline and exhibited a pronounced conchoidal fracture. In addition to uranothorite, thorite is associated with a small percentage of scheelite in the southern Westland material and due to the striking similarity of its optical properties to those of scheelite, it was not recognized at first. However, when the almost non-magnetic fraction, thorite-scheelite-zircon was inspected in ultraviolet radiation of wavelength of approximately 2540 Å., scheelite appeared intensely blue, zircon, gold, whereas uranium-free thorite fluoresced with a faint dull white with a pinkish tinge. If this fraction is treated with aqua regia, washed, dried and then examined under a binocular microscope, scheelite may be readily distinguished from thorite by the distinctive coating of yellow oxide formed on the tungstate.

The following optical properties have been

[27] After removal of more susceptible material uranothorite was attracted in the Frantz separator with a current of 0.9–1.1 amps., cross-slope of 10, and a longitudinal slope of 15.

[*] A monoclinic symmetry has been demonstrated for the uranium-free thorium silicate described herein by recent work (Hutton and Pabst).

TABLE 11.—ANALYSES OF URANOTHORITE

	A	B	C	D	E
PbO	—	0.40	1.32	1.67	—
Pb	0.47	—	—	—	—
UO_2	11.5	—	7.67	—	—
UO_3	—	9.96	9.46	9.78	7.70
ThO_2	62.6	52.07	46.33	50.06	64.56
Ce_2O_3	—	—	0.08	1.39	—
CeO_2	1.1	—	—	—	—
Y_2O_3	—	—	0.36	—	—
FeO	—	—	0.43	—	—
Fe_2O_3	2.3	4.01	0.75	7.60	0.78
FeS_2	—	—	trace	—	—
MnO	—	—	0.07	trace	—
Al_2O_3	—	0.33	0.13	—	—
CaO	0.3	2.34	4.38	1.99	0.95
MgO	0.3	0.04	<0.01	0.28	—
SiO_2	19.5	19.38	19.56	17.04	13.00
H_2O-	0.25	11.31	3.57	9.46	5.30
H_2O+	1.68		5.67		5.74
Insol.	—	—	0.15	—	—
	100.0	99.95	99.94	100.13*	98.03
Sp. Gr. 25° C.	6.25 ± 0.05	4.126	4.414	4.38	—
U	10.14	—	14.63	—	—
Th	55.02	—	40.72	—	—
Pb	0.47	—	1.22	—	—
$\frac{Pb}{(U + 0.36Th)}$	0.0157†	—	0.04	—	—

A. Uranothorite from Gillespie's Beach, Southern Westland; analysis recalculated to 100%, after eliminating zircon, scheelite, ilmenite, and 1.83% of undetermined material. Analyst: F. T. Seelye.

B. Uranothorite from Lake Champlain; total includes alkalies 0.11% (*see* Dana, 1892, p. 488, analysis no. 5).

C. Uranothorite, MacDonald Mine, Hybla, Ontario (Ellsworth, 1927b, p. 370).

D. Uranothorite from Arendal, Southern Norway (*see* Brögger, 1890, p. 121, anal. no. 7).

E. Orangite from basin of Sahavary River, Madagascar (Turner, 1928, p. 76, table 2, analysis no. 16).

* Total includes 0.86% of P_2O_5 and uranium is quoted as U_2O_3.

† This factor multiplied by 7600 gives the age in millions of years as 119. This age determination, which is a minimum figure only since the mineral was not found *in situ*, is equivalent to the lower part of the Cretaceous Period.

determined for this constituent:

Refractive indices:

$\alpha = 1.898 \pm 0.003$

$\beta = 1.900$ (by calculation)

$\gamma = 1.922$

$\gamma - \alpha = 0.024$

Dispersion: moderate with $\rho < v$.

$2V = 25° \pm 2°$

D. at 20°C. = 7.1 ± 0.1 (using approximately 40 mgm. of material).

In a statistical analysis of grains of uranium-free thorite mounted in Canada balsam, 7 per cent of the grains had similar orientation, since such grains exhibited low double refraction and gave interference figures in which the positive acute bisectrix lay just outside the edge of the field of view. Since these grains were anhedral, this orientation suggests cleavage, perhaps parallel to a flat pyramid or basal plane.

A small sample of the present mineral was separated in the pure state by the writer and analysed by Mr. Seelye; only 7 mgm. of material were available for analysis and hence only thorium, uranium, silica, and water determined;

the remainder, determined by difference, was reported to consist of rare earths and Fe_2O_3. For a second pure sample, the writer determined the iron, reported in the ferric state, to be 1.2 per cent, rare earths 2.6 per cent, traces only of manganese and phosphorus, and an

TABLE 12.—ANALYSIS OF URANIUM-FREE THORITE, GILLESPIE'S BEACH

ThO_2	76.6
UO_2	nil
SiO_2	19.7
Fe_2O_3	1.2
CaO	nil
MgO	nil
MnO	trace
P_2O_5	trace
Ce_2O_3 etc.	2.6
H_2O (total)	nil
	100.1

absence of calcium and magnesium. Both uranium and water are absent (Table 12). The chemical composition and the optical and physical properties of the south Westland thorite are quite distinct from those recorded for thorite from numerous other occurrences, and it might be profitable to list the most distinctive features now. These are as follows:

(1) The mineral is colorless. The writer can find no record of colorless thorite; brown, black, or orange-yellow, as in the variety orangite, is usual.

(2) A relatively high birefringence, *viz.* 0.024, is unusual; the mineral is usually reported to be isotropic, and when anisotropic, a birefringence in excess of 0.01 does not appear to be found (Winchell, 1933, p. 185). Brögger (1890, p. 123–124), in thin sections of altered thorite and orangite, has observed what he believed to be strongly birefringent remnants of unaltered thorium silicate, but he did not give exact figures; and Lacroix (1922, p. 243) has listed thorite from Madagascar as uniaxial and positive, but usually isotropic.

(3) The mineral is anhydrous whereas all analyses of thorites listed by Hintze (1915, p. 1675) show a high percentage of water; 18 analyses showed a range from 6.14–12.05 per cent.

(4) The density, determined with approximately 40 mgm. of fine-grained material is 7.1 ± 0.1 at 20°C., greatly exceeding the values previously recorded for thorite. Brögger (1890, p. 120) has stated that the specific gravity of thorite has a range of 4.4–4.8, whereas orangite may range from 5.19–5.40; Dana (1892, p. 481), quoting Chydenius, lists a range of 4.888–5.205 for orangite, and, incidentally, a value as low as 4.126 for uranothorite.

(5) The mineral does not gelatinize upon treatment with a variety of concentrated acids. (*See* Murata, 1943, p. 552.)

(6) The mineral is characteristically biaxial, whereas thorite is stated to be uniaxial, as it should certainly be on account of its symmetry.

These properties suggest that this $ThSiO_4$ is quite distinct from normal tetragonal thorite and that it is some other crystalline modification of $ThSiO_4$. If this is so the mineral should not be called thorite and a new name is required, but the writer is loathe to introduce such a name until further information can be obtained. The lower densities and refractive indices and isotropic character of previously recorded thorites have, in the writer's opinion, resulted solely from hydration of thorite due to metamictization; further, there is no sound basis for assuming that the variety orangite should have a higher density than thorite itself. Several variable factors may affect the physical properties of thorite and its variety orangite, two of the most important being (1) the degree of metamictization, and consequently the extent of hydration, and (2) the extent of substitution by other ions, chiefly uranium and iron.

Doelter's statement (1918, p. 230) makes it clear that he believed that water-content and percentage of uranium in thorites were directly proportional. The few analyses listed by him appear to support that conclusion as do analyses of ferriuranothorite from Vernadsky Mine, Sludianka (Kalinin, 1945, p. 202), and hydrothorite from Wodgina, Western Australia (Simpson, 1928), but the uranothorite from south Westland contains a high percentage of uranium but a low water content. Another decided exception is a ferriferous thorite from northern Kirghizia in which Starik, Kravchenko, and Melikova (1941) found only 0.41 per cent U_3O_8 and 5.64 per cent of water. Since

in thorite, uranium replaces thorium, and it is believed that this substitution does not upset valency relationships, there is little reason to assume a sympathetic relationship between uranium and water content.

A *phosphorus-bearing thorite* occurs as occasional grains in concentrates from Red Jacks, Blackball, and Snowy River dredges, and a sufficient quantity for close study was available only from the two latter sources. In Blackball material, phosphorian thorite was found in the −230 fraction as anhedra with marked conchoidal fracture. The mineral is pale yellow and non-dichroic, uniaxial and positive. The refractive indices and birefringence were determined to be as follows: $\alpha = 1.78$, $\gamma = 1.79$, $\gamma-\alpha = 0.01$. Dispersion, cleavage, and twinning were not seen. In Snowy River dredge concentrates, thorite was restricted to one large grain in screenings coarser than 35-mesh. This fragment was well worn, pale yellow with a density of 6.0. When powdered the mineral was almost completely soluble in concentrated nitric acid. It is quite isotropic and a refractive index of $n = 1.87 \pm 0.01$ was found. Although a very small quantity of this thorite was available a partial analysis (Table 13) shows the presence of 4 per cent of P_2O_5. Careful microscopic examination of the powdered mineral left no doubt that it was quite homogeneous and not a mechanical mixture or aggregate of two or more constituents. Chemically the mineral is similar to auerlite (Table 13, B and C) in which considerable substitution of $(SiO_4)^{4-}$ groups by $(PO_4)^{3-}$ is believed to have occurred. Such replacement is not uncommon and has been clearly shown to occur in orthite-nagatelite (Machatschki, 1931), griphite (McConnell, 1942), zircon (Hata, 1938b), and garnet (Mason and Berggren, 1941). Indeed, a theoretical consideration of ionic sizes and radius ratios would allow prediction of such an interchange since the Si–O distance of 1.62 Å., is exceedingly close to the P–O distance in a structure such as xenotime, 1.64 Å. Simultaneous compensation for the electrical neutrality could be accomplished for phosphorian thorite by replacement of Th^{4+} by Fe^{3+} or Al^{3+}, or, in agreement with the composition of the New Zealand phosphorian thorite, by the trivalent rare earths, Ce^{3+}, Y^{3+}, etc. Although substitution of phosphorus for silicon is even more extensive in auerlite, analyses B and C, compensatory replacements do not appear to be adequate. However this disagreement may be more apparent than real since both B and C are old analyses,

TABLE 13.—ANALYSES OF PHOSPHORIAN THORITES

	A	B	C
ThO_2	74	72.16	70.13
SiO_2	16	6.84	7.64
P_2O_5	4	8.58	7.46
TiO_2	1	—	—
Rare earths	3	—	—
Fe_2O_3	0.2	1.78	1.38
Al_2O_3	nil	—	1.10
CaO	nil	—	0.49
MgO	nil	—	0.29
H_2O	2	10.64	11.21
	100.2	100.00	99.70

A. Phosphorian thorite from Snowy River. U, Zr, and V absent. Determination of rare earths, iron, and water by C. Osborne Hutton, remainder by F. T. Seelye.

B. and C. Analyses of so-called auerlite from North Carolina. *Vide* Doelter, 1918, p. 234.

and in analysis B the thorium figure was obtained by difference only, presumably on the supposition that the constituents actually determined would be the only others present. In view of the hypothesis that extensive substitution of rare earths and phosphorus for thorium and silicon respectively occurs in thorites, it is questionable whether the name auerlite need be retained for a mineral that is merely a member of an ionic substitution series. The same argument certainly applies to hydrothorite (Simpson, 1928a, p. 464–465; 1928b) which appears to be only a metamict thorite containng a little more water than usual and some substitution by phosphorus, CaO, and rare earths as in auerlite.

Considering the full extent to which such substitution could occur in thorite, some of the relevant facts are: (1) The similarity of the dimensions of the interionic distances P–O and Si–O. (2) The comparable sizes of the ionic radii of Th^{4+} and Y^{3+}, 0.95 Å (Zachariasen's value for 6-fold coordination) and 1.06 Å.

With complete replacement of Th^{4+} and Si^{4+} by Y^{3+} and P^{5+} respectively, it is clear that we

have the mineral xenotime, YPO_4, and clearly the phosphorian thorite described herein appears to be an intermediate member of an isomorphous series thorite-xenotime. Whether this isomorphous replacement is unlimited or restricted within narrow limits is not known at present and an answer to this question must await further work. From a structural point of view, complete substitution should be possible. However, many of the analyses of xenotime listed (Doelter, 1918; Hintze, 1922) show appreciable but limited quantities of silica. A number of these determinations are old or incomplete but, in the best of these, appreciable silica parallels the incoming of a noteworthy amount of thorium or uranium, or both. More recently Hata (1938b) has published a more than usually complete analysis of pure xenotime in which he found 4.32 per cent SiO_2, and 2.47 per cent ThO_2, and incidentally, 1.90 per cent ZrO_2 also. Substitution of phosphorus by silicon is not necessarily the only interchange that may occur in xenotime, for undoubtedly as in hussakite (Kraus and Reitinger, 1901), some substitution of $(PO_4)^{3-}$ by $(SO_4)^{3-}$ groups is possible, apparently compensated for by some interchange of calcium for yttrium earths. Replacement of $(SiO_4)^{4-}$ by $(PO_4)^{3-}$ does occur in thorites, but its extent must await further study. Since this interchangeability is possible, it will be of interest to know whether $(AsO_4)^{3-}$ groups are also present in some thorites; most noteworthy in this respect is the listing by Starik *et al.* (1941) of 0.51 per cent As_2O_5 in ferrithorite. Actually arsenic was probably never looked for in the analysis of thorites; almost certainly neither was phosphorus in older work.

TITANITE: Titanite is a constituent of most of the dredge concentrates but in no instance makes up more than 3 per cent of any particular fraction of a sample. However, pay-streaks of titanite were found in many of the beach sands east of Tarakohe Inlet, in northwest Nelson, and in some of these titanite constitutes up to 50 per cent of the sand and is associated chiefly with zircon, apatite, ilmenite, and magnetite.

In the dredge concentrates, titanite is usually limited to the finer grade-sizes and exhibits a range of color from pale yellow to light brown. In most samples anhedral grains were found, with spindle-shaped and double-wedge-shaped crystals less common. In the north-west Nelson beach sands, however, sharply euhedral titanite crystals are abundant (Pl. 2, fig. 38; Pl. 3, figs. 39–42). They are fairly simple in form and the dominant habit exhibits a combination of *l* (112), *c* (001), and *x* (102) with the latter form usually predominant; actually, the crystals are very similar in appearance and general habit to those described and figured by Webb (1939, p. 345). As a result of this habit, crystals tend to orient themselves with the flat dome in the plane of the microscope stage when immersed in oils, and, since the positive acute bisectrix Z is nearly normal to that plane, crystals showing good centered interference figures with characteristic dispersion are usually readily obtained. Cleavage, probably parallel to the unit prism, was observed in fractured crystals. Titanite from this area is pale yellow or very nearly colorless in grains with diameters less than 62 microns.

In the $-35 + 60$ fraction of a beach concentrate recently sent the writer from Bruce Bay, South Westland, crystals of titanite of unusual appearance were found with the following properties:

(1) The crystals are black, euhedral, and exhibited the simple "envelope" habit.

(2) In transmitted light the crystals are slightly translucent at the edges but the dull-black opaque appearance in oblique illumination is clearly due to dense clouds of dustlike particles.

(3) The magnetic susceptibility of the crystals is no greater than the associated but perfectly clear, inclusion-free titanites.

(4) Chemical tests showed only traces of Fe^{3+} and Mn^{3+}, in addition to the usual constituents of titanite.

(5) After crystals were heated at dull-red heat for 2–3 minutes, they became more translucent.

In view of the absence of any increase in magnetic susceptibility above that usually found in titanite, the rarity of Fe^{3+}, and the tendency for the opacity to clear up on heating, it is believed that the inclusions are carbonaceous, rather than an iron oxide dust.

The following optical properties for titanite from Wainui Inlet, northwest Nelson were determined in sodium light: $\alpha = 1.912 \pm 0.003$; $\beta = 1.917$; $\gamma = 2.08$; $\gamma - \alpha = 0.168$; $2V = 36° \pm 2°$. Pleochroism, faint but more distinct in

crystals with diameters greater than 0.25 mm., followed the scheme: X = pale golden-yellow; Z = slightly deeper golden-yellow; Z > X. The density, determined by centrifuging the powdered mineral in Clerici solution, was 3.54 ± 0.02.

Zachariasen's (1930) work on the crystal structure of titanite has shown quite clearly that the sum of the valencies of the bonds going to the three differently positioned oxygen atoms is not in good agreement with Pauling's second rule as understood for ionic crystals, but that there could be better agreement if one of these oxygen positions is replaced by hydroxyl; in a footnote Zachariasen suggests that fluorine might possibly replace hydroxyl in some titanites (1930, p. 15). Two other investigators, Bellanca (1942) and Morgante (1943) have also suggested the presence of the group (O, OH, F) in titanite; Bellanca merely included such a group in his suggested formula, whereas Morgante actually determined the presence of (OH) analytically. In this connection it is interesting to observe that Strunz (1937, p. 10–12) believed that titanite and tilasite were isotypic, tilasite having one atom of fluorine and the formula $[(AsO_4)F]$ MgCa. In view of these conclusions, and Sahama's (1946) later determinations, a pure sample of northwest Nelson titanite was prepared for analysis by fractionating the graded sand electromagnetically[28], and subsequently centrifuging the powdered minineral in Clerici solution; both hydroxyl and fluorine ions were particularly looked for. In addition, since crystals of this particular titanite produced radiographic imprints on a photographic plate, the presence of both thorium and uranium were looked for. The analysis (Table 14, anal. A) shows 0.67 per cent of fluorine whereas the loss on ignition amounted to 0.93 per cent. Unfortunately, determination of structural water by this method is not reliable since Sahama (1946, p. 103–104) has shown conclusively that an indefinite quantity of fluorine could be lost. The total quantity of rare earths amounts to the relatively high figure of 1.06 per cent, not sufficiently high to warrant classifying the mineral as yttrotitanite or keilhauite. On the other hand since both alumina and ferric oxide are high, the northwest Nelson titanite would best be classified as the variety grothite (Doelter, 1918, p. 61–62). Uranium was absent but the 0.28 per cent of ThO_2 found explains the radioactivity of the mineral. The occurrence of radioactivity in titanite is not uncommon, as Larsen and Keevil (1942, p. 209) and other workers have already shown, and some degree of radioactivity can be expected since the Ca, U and Th ions are not greatly dissimilar in size. A richly uranoan or thorian titanite is not to be expected structurally.

Zachariasen (1930, p. 15) gives the formula of titanite as $CaTiSiO_5$ but admits that, since most analyses show water to be present, presumably as (OH), a formula such as $ABSiO_4X$,

TABLE 14.—ANALYSES OF TITANITE

	A	B	C
SiO_2	30.35	30.32	30.48
TiO_2	35.44	33.70	38.00
CaO	26.46	28.40	27.32
Fe_2O_3	2.50	nil	nil
Al_2O_3	2.15	4.44	0.05
MnO	0.25	0.01	0.08
MgO	0.10	0.16	nil
K_2O	0.02	nil	nil
SrO	nt.fd.	0.005	0.29
ThO_2	0.28	nt.dt.	nt.dt.
Ce_2O_3	0.03		
La_2O_3 group	0.85	0.30	0.26
Y_2O_3 group	0.18		
F	0.67	1.40	0.68
Loss on ignition	0.93	0.54	0.61
H_2O at 105°C.	0.17	0.18	0.12
	100.38	100.61	100.63
O for F	0.28	0.59	0.29
	100.10	100.02	100.34

A. Titanite from Wainui Inlet, N. W. Nelson. Analyst: F. T. Seelye.

B. Grothite from Parainem, Finland. Analysts: Th. G. Sahama *et al.* N. B. total includes Nb_2O_5 0.69, Ta_2O_5 0.03, V_2O_5 0.09, Cr_2O_3 0.02, FeO 0.20, ZrO_2 0.06, BaO 0.003, Na_2O 0.16. (Sahama, 1946, p. 94, anal. No. 3).

C. Titanite from Kola Peninsula, Russia. Analysts: Th. G. Sahama *et al.* N. B. total includes Nb_2O_5 1.52, Ta_2O_5 0.02, V_2O_5 0.005, FeO 0.10, ZrO_2 0.05, BaO 0.006, Na_2O 1.05. (Sahama, 1946, anal. No. 5.

[28] North-west Nelson titanite is slightly paramagnetic and therefore it was possible to use electromagnetic means for concentration. Saporte and Canal (1948, p. 21) list titanite as having a lower magnetic susceptibility than zircon, but this is not generally the case.

in which X represents either O or (OH) or both, would seem more appropriate. Morgante (1943), Sahama (1946, p. 106), and Jaffe[29] (1947, p. 641) all suggest a formula almost identical to

TABLE 15.—RECALCULATION OF ANALYSIS OF NORTH-WEST NELSON TITANITE

	No. of (O·OH.F)	No. of metal atoms		
SiO_2	1.0116	3.94		4.00
Al_2O_3	0.0633	0.33	0.06	
			0.27	
TiO_2	0.8872	3.45		3.98
Fe_2O_3	0.0468	0.24		
MgO	0.0025	0.02		
CaO	0.4725	3.68		
La_2O_3	0.0078	0.04		
Y_2O_3	0.0027	0.01		3.77
MnO	0.0035	0.03		
ThO_2	0.0036	0.01		
F	0.0352	0.27		1.07
(OH)	0.0516	0.80		

Formula: $(Ca, La, Yt, Mn, Th)_{3.77}(Ti, Al, Fe, Mg)_{1.03}[(SiAl)_4O_{19}](OHF)_{1.07}$

Zachariasen's alternative, and Morgante expressed his understanding of the possible substitutions as follows:

Formula: $ABSiO_4X$ where
A = Ca and cations with radii 0.67–1.35 Å.
B = Ti and cations with radii 0.57–0.67 Å.
and X = O, OH, F.

The analysis of Nelson titanite has been recalculated (Table 15) on the basis of 20 oxygen atoms to the unit cell and the figure obtained for "loss on ignition" has been interpreted here as structural water, although this figure may not be absolutely reliable. Calcium is replaced to some extent by rare earths, manganese, and thorium, and the valency upset resulting from these substitutions is made good with replacements of titanium by aluminum, trivalent iron, and magnesium. To make up the silicon group to the theoretical figure of 4, a small quantity of aluminum, such as 0.06, is considered to have coordination number 4; this is to be expected because of its radius ratio with oxygen of 0.41 (Pauling, 1948, p. 382). This substitution leaves the titanium group negligibly deficient, but the figure for the calcium group containing large ions is very much lower than the value required structurally. On the other hand the (O, OH) group, in spite of the method of determination, is high compared to the figures given by Sahama (1946, p. 96–97), and accordingly this Nelson grothite seems to show a greater degree of replacement of the O_1 positions by (OH) and F than is to be seen in other modern analyses available.

Thus the following formula expresses the mutual substitutions found in Nelson grothite: $(Ca,La,Y,Mn,Th)_{3.7}(Ti,Al,Fe''',Mg)_{3.9}(SiAl)_4(O,OH,F)_{20}$.

TOPAZ: Topaz was recognized in sands from Okarito, Grey River, Gillespie's Beach, and Kaniere River dredges but it did not exceed 1 per cent of any of the fractions therefrom. The mineral was found as colorless, glassy, water-clear grains of tabular or platy habit due to the dominance of the (001) cleavage; a roughly square-shaped outline was not uncommon, otherwise fragments were irregular with prominent conchoidal fracture (Pl. 3, figs. 43–45). Strings of minute, colorless globular inclusions of very much lower refractive index than topaz, but otherwise unknown character, were present in occasional crystals.

The following physical properties were determined for topaz from Okarito: $\alpha = 1.623 \pm 0.002$; $\beta = 1.626$; $\gamma = 1.632$; $\gamma - \alpha = 0.009$; $2V = 50° \pm 2°$; $D. = 3.52 \pm 0.02$.

These data suggest that the Okarito topaz has a relatively low fluorine content (Penfield and Minor, 1894, p. 393) but with less hydroxyl than the low-fluorine topaz described by Pardee *et al.* (1937, p. 1064).

TOURMALINE: It was surprising to find tourmaline represented sparsely among the mineral assemblages studied here, since many of them were derived from distributive provinances in which granites are abundant. In no instance did tourmaline exceed 2 per cent of any fraction and in many only an occasional grain was found; the mineral was absent altogether from some dredge concentrates and most beach sands, although there was a tendency for it to be a more important constituent of panned concentrates.

[29] Presumably Jaffe's use of two molecules of silicon in the formulae given on p. 641 is merely a typographic error.

Where present, tourmaline was restricted in the main to the finer fractions. The most common mode of occurrence was as slender to short stumpy prismatic crystals that frequently showed terminal forms at one end only; in many crystals there were distinctive cross-fractures, otherwise in irregular fragments with subconchoidal fractures. Normally the characteristic dichroism with maximum absorption perpendicular to the length of the prismatic crystals is an important diagnostic feature but care must be used in this connection. In several assemblages tourmaline fragments were found with slender prismatic form but with the elongation parallel to the Z vibration direction; therefore, these fragments showed maximum absorption parallel to the major axis. Such forms may readily be developed in the first place from stout prismatic crystals of tourmaline by the appropriate fracturing of those crystals along closely spaced planes of basal parting. This point is stressed here because cursory examinations of preparations containing such fragments might result in confusion of this mineral with some of the blue sodic amphiboles of low extinction angle. Perhaps it would be pertinent to stress a further point at this stage, *viz.* that fragments with shapes mainly determined by closely spaced (001) fractures may be tabular in that direction and therefore show a preferred orientation in the mounting media. Such fragments are isotropic but no confusion in their identification need arise if reasonable care is taken in determination.

The tourmalines found in the various sands showed a wide range of color, and dichroic schemes of the most characteristic of these are tabulated below in Table 16. Those detailed in Table 16 are not necessarily restricted to the localities mentioned except in a few instances, such as the green tourmaline in Kaniere material and the black variety in the beach sands of Wainui Inlet.

Fine dust-like clouds of opaque iron oxides are frequently included within the brown, brownish-purple and smoky-grey varieties which in this respect closely resemble the typical "schist" tourmalines described by Hutton and Turner (1936, p. 264, 268). In some sands, particularly those of Lowburn, in Central Otago, multi-colored tourmalines predominate in which one extremity is colored differently, often a deep brown, from the remainder of the crystal, which is commonly purple or purplish-brown (Pl. 3, figs. 55–56). Most of the tourmalines found in the material used in this investigation appear to belong to the dravite-schorlite series with the refractive index for the ordinary ray showing a range of 1.639–1.664, but the varieties in which pink or green tints predominate may possibly lie closer to the schorlite-elbaite solution series. Further work on the tourmalines in these sands is in progress.

TABLE 16.—DICHROIC SCHEMES OF TOURMALINE

Locality	E	O
Barrytown	pale yellow	deep brown
Barrytown	colorless	brownish-green
Barrytown	pale yellow	deep purplish-brown
Arahura	pale pink	deep purple
Kaniere	colorless	green
Kaniere	pale yellow	greenish-brown
Lowburn	colorless or nearly so	brown, blue, or smoky grey
Wainui Inlet	brown	black

XENOTIME: The yttrium phosphate, xenotime (Pl. 4, figs. 64–68), has a similar distribution to that of monazite in the samples investigated in this study, but the former mineral was very much less plentiful. Its mode of occurrence, physical and chemical properties have been described in an earlier paper (Hutton, 1947a) and need not be repeated here. However, further determinative work has found the following range of refractive indices: O = 1.720–1.724 ± 0.003; E = 1.820–1.828; E − O = 0.101–0.107. These refractive indices are distinctly higher than any hitherto recorded for xenotime so far as the writer is aware, and since, owing to the rarity of the material, quantitative chemical data are not available, these optical properties cannot be correlated therewith. However, because isolated crystals of xenotime from the Grey River area have been shown to affect an alpha-ray plate, they must contain either thorium or uranium or both and the high values for the refractive indices can reasonably be attributed to the presence of either or both of these ions. Although the exact composition of the xenotime described here is unknown, it is pertinent to point out that any replacement of

yttrium by uranium or thorium must be accompanied by complementary substitution of phosphorus by an ion of lower valency, most likely silicon. Replacement of $(PO_4)^{3-}$ by $(SiO_4)^{4-}$ is unlikely to affect the refractive indices of the mineral to any significant degree, and since the ions of phosphorus and silicon have similar radii, the substitution should not cause significant density changes. Therefore, determination of refractive indices and density would probably give a close estimation of the cumulative percentages of uranium and thorium in xenotime.

Xenotime can be distinguished from yellow zircon, monazite, or titanite relatively simply by optical means alone (Hutton, 1947a) but bastnäsite, a fluocarbonate of the cerium metals may be confused with xenotime. Both minerals have such similar optical properties that diagnosis on that basis alone would be doubtful. Geijer (1920, p. 11) gave O = 1.7225 and E′ = 1.8242 for bastnäsite from Bastnäs, and Glass and Smalley (1945, p. 609) record almost identical figures for material from Belgian Congo; for other occurrences, the refractive indices were found to be very slightly but not significantly lower. Distinction on the basis of density is no more helpful; although Winchell (1933, p. 85) lists the density of bastnäsite as 5.0, distinctly higher than that usually recorded for xenotime, Silberminz (1929) has determined the density of analysed bastnäsite from Kyshtymsk District to be 4.746. While working with powdered bastnäsite from Gallinas Mountains, Lincoln County, New Mexico, the writer found that in the absence of analyses or spectrographic data, the presence of a perfect cleavage and its solubility with effervescence in warm HCl acid conveniently served to distinguish it from xenotime. The latter test worked satisfactorily with −230-mesh material, and particularly so if the grains are warmed on a glass slide before a drop of acid is added with a fine dropper.

Zircon: The mineral zircon is ubiquitous in all specimens containing material of average grain-size less than 0.25 mm. It usually reaches its highest frequency in the −230-mesh screenings, but in almost all instances the frequency declines so rapidly in the coarser grade-sizes that the mineral may almost fail in a number of samples in −60 +120 screenings; it is generally absent from the coarsest grade-sizes.

Ordinary colorless normal zircon (Pl. 1, figs. 1–9) and the pale-pink to deep-purple variety, hyacinth (Pl. 1, figs. 10–19; Pl. 4, fig. 60), have widespread distribution.

The term *"normal"-zircon* is used here in the sense employed by Anderson (1941) to distinguish normal zircon of high density, refractive index, and birefringence from malacon or *"low"-zircon*. The term "normal"-zircon is considered synonymous with "high"-zircon employed by Smith (1940, p. 280). Stevanovic (1903, p. 247–252) was probably first to suggest subdivision of the zircon group on the basis of density, optical properties, and their reaction to heating. He divided zircon into three categories: (1) zircon of density about 4, (2) zircon of density about 4.7, and (3) zircon of density between those of (1) and (2). Stevanovic reported that on heating, low-density zircon exhibited no change in properties but that the refractive indices of zircon of intermediate density did change. These varieties of zircon have been called α-, β-, and γ-zircons (Spencer, 1904, p. 47–48; Smith, 1940, p. 280; Wadia and Fernando, 1944, p. 36–7; compare, however, Schlossmacher, 1932, p. 562). Chudoba and Stackelberg (1936, p. 243), following this general classification, showed that both α- and γ-zircons would change into β- or "high"-zircon after heating to 1450°C. for 6-hour periods.

Every gradation from sharply euhedral crystals exhibiting a considerable variety of forms and habit, through slightly abraded crystals to well-worn anhedra devoid of any crystal faces were found in most samples. In addition to these, irregularly shaped fragments, flakes, or splinters with pronounced conchoidal fracture are not uncommon. Generally speaking, however, a higher percentage of unworn euhedra were to be found in the finest grade sizes, whereas zircon attained a greater degree of rounding as the grain-size became progressively coarser; when zircon was present in fractions with average grain size greater than 0.5 mm. the particles were predominantly anhedral. Among the coarsest material from one sample of concentrates from Slab Hut dredge, −18 +35 mesh, a few almost perfectly polished anhedra of zircon were obtained. These grains showed no remnants of crystal faces, were slightly flattened and ellipsoidal in outline with the major and minor axes approximately 0.85

mm. x 0.50 mm. In the three particles examined, the major axis was not parallel to the vertical crystallographic axis, obliquity being about 12°, 16°, and 21°. Since it is difficult to visualize the process whereby unbroken crystals of zircon would be abraded and polished naturally into forms having directions of elongation other than those parallel to the *c* axis, one can only assume that these grains represent the final product of abrasion of splinters of zircon crystals of considerably greater size than any so far found in any West Coast sands studied. These grains of zircon were water-clear, very faint yellow, and of gem quality.

The most usual habit found in normal zircon showed a development of second-order prisms *a*(100) terminated by first-order pyramids *p*(111) with or without third-order pyramidal faces; *m*(110) is usually present but rarely dominant (Pl. 1, figs. 8–9). Less often, second-order prisms are terminated by three orders of pyramids. Very occasionally a combination of *m*(110) and *p*(111) only was found (Pl. 1, fig. 2) but in general crystals with this simple habit are very rare in any of the New Zealand occurrences known to the writer, except in the case of hyacinth described later; however, this habit appears to be fairly common in localities elsewhere (the Dartmoor Granites, Brammall and Harwood, 1923, p. 27) although not found for zircons from Falmouth Granite in Cornwall by Ghosh (1928, p. 333). Crystals are often considerably distorted so that a crystal with *a*(100) well developed and *m*(110) poorly developed may be terminated by large first-order and small second-order pyramids at one extremity, whereas at the other end of the crystal the degree of development of these pyramidal faces is reversed, and at the same time a small basal plane is sometimes present. The basal plane *c*(001), however, is uncommon. In sands from Snowy River, a small percentage of a zircon of distinctive habit was noted. It differed from those previously described on account of the predominance of steep pyramids, probably *u*(331); and whereas the ratio length to breadth in zircons from other localities was usually approximately 2:1, the distinctive Snowy River zircons had a ratio of approximately 3:1, and occasionally 4:1 (Pl. 4, figs. 61–62). Except in Snowy River sands, and the rare crystals in material from Grey River and Slab Hut dredges, this particular habit was not found in any other material studied.

Inclusions are frequent and varied, and only slight success resulted from attempts to separate and determine them. Selective solution of finely crushed zircon in warm hydrofluoric and sulphuric acids generally failed because the inclusions were also soluble in acid, and separation of foreign particles from finely crushed zircon centrifugally was not satisfactory in every instance because it was only convenient to use Clerici solution and zircon as well as most of the impurities had greater density than the liquid, that is 4.2. However the following inclusions have been noted:

(1) Spherical and sausage-shaped cavities, some probably gas-filled, though in rare instances globules of fluid partially filled the internal cavities.

(2) Magnetite and ilmenite.

(3) Euhedral zircon.

(4) Pale-yellow to nearly colorless monazite and xenotime; these were often seen in zircons obtained from Grey River sands.

(5) Slender, acicular, colorless crystals with refractive index close to 1.65 and relatively insoluble in acids; possibly mullite but not proven.

(6) Dark-brown, prismatic grains of rutile.

(7) Reddish-brown, isotropic grains of cubic or tetragonal aspect soluble in hydrofluoric and sulphuric acids, the solution of which gave reactions for tantalum and niobium; they might be any one of a number of brown tantalian minerals such as samarskite, euxenite, polycrase, pyrochlore, fergusonite, yttrotantalite in the metamict state. Their nature was not verified, owing to insufficient material.

(8) Clouds or stringers of fine black dust.

(9) Subhedral particles of pale-brown, poorly dichroic cassiterite.

(10) Short, stumpy prisms of brownish tourmaline.

Any orderly arrangement or preferred orientation of inclusions was most unusual, although the presumably gas-filled cavities are in some instances elongated parallel to the vertical crystallographic axis.

In some of the samples studied, the constituents have been stained by iron solutions so that the zircons have a distinctly yellow color. The ferruginous stain was very evenly deposited on the crystal surfaces and was not removed by the abrasive action of panning or shaking on concentrating tables. With a hand lens, and in

some instances even when immersed in oils and examined under the microscope, these yellow-stained zircons had the appearance of monazite; and it should be stressed that this coloration has misled some previous workers to believe that they were dealing with monazite-bearing sands. Actually, such stains are readily removed by treating the material concerned with warm 1:1 hydrochloric acid for 5 or 10 minutes, and then washing carefully with copious warm water to which a few drops of hydrochloric acid has been added.

Zoning has been observed in a few cases (Ngahere and Kaniere dredges) but it is generally uncommon in normal zircon; it consisted of numerous exceedingly fine, almost thread-like concentric bands in a number of crystals; these zones, which alternated with the host zircon, seemed more prominent nearer the periphery than towards the center. Generally the fine zonary banding could only be observed when the vertical crystallographic axis of a zoned crystal was placed parallel to the vibration direction of the lower nicol. In this position the refractive index of the fine bands was found to be appreciably less than the appropriate refractive index of the adjoining zircon. When, however, such a crystal was rotated through 90° so that the vertical crystallographic axis was perpendicular to the plane of polarized light, little or no zoning was apparent. It would seem then that, as the zircon crystals grew, some oscillation of the composition of the supplying fluid must have occurred, resulting in abrupt changes in the composition of the crystallizing zircon such that the refractive index for the ordinary ray was similar in all zones but the refractive index for the extraordinary ray showed a distinct range of values.

Twinned crystals of zircon (Kaniere River) exhibited the usual knee-shaped forms resulting from twinning about (101).

Refractive indices of a number of typical normal zircons were measured in order to observe whether metamictization had occurred to any extent. However, these figures (Table 17) do not suggest any significant alteration or isotropization, and the range of values obtained lie within the limits listed by Winchell (1933, p. 184) for that mineral, though distinctly higher in a few instances than the values determined by Morgan and Auer (1941, p. 308). No chemical analyses of normal zircon have been made, but numerous tests with a Geiger-Muller counter indicate an exceedingly low percentage of uranium or thorium or both. For instance, whereas 1 per cent of U_3O_8 gives ionisation current of 16,800 amperes x 10^{13}/sq. cms., Naseby zircon gave a value of 8 and Lowburn zircon only 7.

TABLE 17.—REFRACTIVE INDICES AND SPECIFIC GRAVITY OF SOME NEW ZEALAND ZIRCONS

	Naseby	Lowburn	Ngahere	Arahura
O	1.935–1.938	1.935–1.939	1.923–1.927	1.934
E	1.978–1.982	1.979–1.984	1.982–1.984	1.989
E–O	0.042–0.047	0.043–0.048	0.055–0.059	0.055
D	4.64–4.67	4.65–4.67	4.62–4.68	4.65

Very rare crystals of green, *phosphorian zircon* were isolated from the −120+230 mesh screening of one of the concentrates from Snowy River. The morphological development of these crystals differed in no important detail from those of normal zircon just described, but the refractive indices and density are decidedly lower, as follows: O = 1.921; E = 1.975; E − O = 0.054; D. = 4.56; Color: = Ridgway's Rivage green, 31′ Y − G. These properties suggest comparison with olive-green beccarite from Ceylon, except that the New Zealand mineral is uniaxial. This green zircon has no greater degree of radioactivity than the colorless varieties referred to and its magnetic susceptibility is similar. Although insufficient material was available for quantitative analysis, some qualitative tests showed that, in addition to zirconium and silica, both phosphorus and rare earths were present, but no hydroxyl. Thus, although the refractive indices and density are slightly lower than those recorded for normal zircon, the absence of hydroxyl suggests that the mineral was not altered in any degree to a

metamict condition. On the other hand, the presence of phosphorus and rare earths suggests some diadochic substitution of silicon and zirconium. As previously seen, ionic substitution of silicon by phosphorus is found in a number of minerals on account of similar ionic size and suitable radius ratio and, in that case, the upset in electrical neutrality can be met by substitution elsewhere in the structure. Thus in this instance, from the evidence obtained by qualitative determinations alone, we assume a replacement such as $Si\ Zr \rightleftharpoons P\ R$, where R = Y, Ce, La etc. Several instances of green and brown zircon containing rare earths, with or without phosphorus, have been found by other investigators and a summary of the appropriate data might be relevant here. Naegite from Naegi, Mino Province, Japan (Shibata and Kimura, 1923b, p. 3–5) contained 6.68 per cent of rare earths in addition to thorium and uranium, but no phosphorus appears to have been found therein, whereas a very similar mineral, hagatalite from Hagata, Iyo Province (Kimura, 1925, p. 82–84) contained 13.1 per cent of rare earths, and again with uranium and thorium but apparently no phosphorus. However, oyamalite, also from Iyo Province, has been shown (Kimura, 1925, p. 84–85) to contain 7.6 per cent of P_2O_5 in addition to 17.7 per cent of rare earths. Still later Kimura and Hironaka (1936) and Hata (1938b, p. 620–622) described as yamagutilite zircon that obviously did not differ in any essential detail from the oyamalite variety of zircon; in those specimens of yamagutilite rare earths amounted to 15.89 and 10.93 per cent and P_2O_5 to 4.23 and 5.30 per cent respectively. All of these varieties are probably metamict, for they contain notable percentages of water. While substitution of silicon by phosphorus and zirconium by rare earths is quite understandable, the presence of a considerable percentage of rare earths without some compensatory replacement of other ions as in naegite and hagatalite requires closer study. Two possibilities are: (1) The minerals may be impure and the rare earths contained in inclusions of samarskite, fergusonite etc. (niobium and tantalum are listed in both analyses, and in addition both minerals are associated with fergusonite in pegmatites). (2) Phosphorus may be present but not recognized.

Whatever the answer, the separate names oyamalite and yamagutilite are unnecessary since these minerals are obviously varieties of zircon and following Schaller (1930) could be termed phosphorian zircons. On the basis of the author's qualitative determinations and by comparison with these Japanese minerals, the Snowy River material is termed phosphorian zircon.

The term *hyacinth* is used here to denote pink, purple, or reddish zircon as distinct from isotropic malacon of relatively low refractive indices and density. It was a constituent of every dredge sample studied although most abundant in material from the Grey River; it is found in most size fractions.

Although no quantitative chemical determinations have been made, the Geiger-Muller counter indicated that the hyacinths are distinctly more radioactive than were the colorless normal zircons; an average reading of 45 for hyacinth as compared with 7 for colorless zircon. Both were unattracted by a magnetic field of high intensity when free from inclusions.

The hyacinth showed a distinct range from very pale pink through pink to a deep purple and, in occasional crystals, purplish-brown. Crystals are generally clear and transparent, although a number were somewhat turbid on account of aggregates of dust-like inclusions or networks of cracks. The color is best seen when the crystals are immersed in liquid with refractive index close to that of the mineral itself; ordinarily the very high refractive index of hyacinth causes excessive internal reflection when examined in the usual mounting media. Distinction between colorless zircon and pale hyacinth is most pronounced, however, when the mineral assemblage containing them is mounted on a white card, or better a vitrosil plate, and viewed in oblique illumination beneath a binocular microscope. Hyacinth differs from colorless zircon also as follows:

(1) The average grain-size is greater. Comparing the colorless zircon-hyacinth ratios in the −230 and −60 +120 mesh screenings, the frequency of zircon relative to hyacinth is found to decrease with increase in grain size.

(2) Whereas colorless zircon has been shown to exhibit a range of grain shapes, from strictly euhedral crystals to abraded and moderately rounded forms, euhedra were much more frequently found; instead, euhedra of hyacinth are

rare (Pl. 1, fig. 10) and the mineral is more frequently found as well rounded often nearly spherical grains (Pl. 1, figs. 18–19).

(3) In grains that exhibited little or no abrasive effects (Pl. 1, fig. 10), the habit of the crystals was almost always simple; a combination of $m(110)$ and $p(111)$, with occasionally a development of $u(331)$, was most usual.

(4) Zonary structures are common (Pl. 1, figs. 11, 12, 15).

(5) A curious system of radiating or anastomosing cracks was seen in a number of hyacinth crystals (Pl. 1, fig. 13); nothing comparable has been found in colorless zircon.

(6) Outgrowths, as distinct from zonary structures, of purple zircon on hyacinth (Pl. 1, fig. 14), are rare; secondary outgrowths have not been observed in colorless zircon.

It is not necessary to enlarge upon the characteristics mentioned under (1)–(3) above, but the remaining observations require some elaboration. A number of distinct zonary patterns are apparent from a glance at Plate 1, but in many examples the very fine concentric banding (Pl. 1, fig. 15) like that described for colorless zircon, is distinct and clear when the vertical crystallographic axis is parallel to the plane of polarized light but is usually invisible or nearly so when the microscope stage is rotated through 90°. Actually it is similar to the type of zoning figured by Bruce and Jewitt (1936, p. 202, text-fig. 5, nos. 20, 23) in colored zircons from some Canadian Precambrian granites, and described by Brammall (1928, p. 28) in purple zircons from Dartmoor. Other zonary structures are:

(1) A unilateral arrangement in which purple zircon appears to have had two wide and successive layers of similar material added, but at one end only (Pl. 1, fig. 11); in some instances only one such zone was present.

(2) Later crystallization of hyacinth partially surrounding earlier but sharply euhedral purple zircon (Pl. 1, fig. 12).

(3) Occasional grains consisting of a later development of hyacinth around well rounded hyacinth and in optical continuity therewith; the later-formed zircon itself may sometimes exhibit very fine concentric zonary development (Pl. 1, fig. 15).

(4) And finally, and probably most common of all, crystals with deeply colored nuclear or central areas that show either an abrupt or a gradational transition towards paler peripheral zones (Pl. 1, fig. 13). In many examples the dark central portions of such crystals are turbid, as if charged with clouds of exceedingly fine dust-like particles. One crystal that may be included in this category was found to have a central zone of low birefringence in comparison with the peripheral portion of the grain; this suggests a transition of the central zone towards a malacon-like material. No crystals were found to contain true isotropic malacon, however.

The systems of irregular cracks developed within a number of hyacinth crystals are quite distinctive and in the majority of examples seem to originate at, or radiate away from, central dark-brown or opaque inclusions. In this respect these zircons are very similar to those described by Robertson (1923, p. 493) from alluvial material of western Togoland. Further, the irregular micro-fissuring is comparable in most respects to that development in the uranothorite crystals and, as pointed out earlier, such characteristics are frequently found in or associated with radioactive minerals.

Secondary outgrowths upon hyacinth were only occasionally found in the several hundred preparations. In the best example (Pl. 1, fig. 14) pale-purple hyacinth, verified as zircon by refractive index determinations, has grown on one of the prism faces of a broken but otherwise euhedral hyacinth so that the vertical crystallographic axis of the sub-individual is normal to the direction of that axis in the host crystal; the secondary purple zircon is clear, unzoned, free from inclusions, and has slightly higher refractive indices than those of the host. These data are as follows:

	Host grain	*Outgrowth of zircon*
O =	1.921	1.929
E =	1.969	1.978

In a number of examples refractive index determinations proved that the outgrowth material was zircon. Therefore, at least for the material from the Grey River, Brammall's suggestion (Smithson, 1941, p. 104) that the secondary material might be some rare earth mineral isomorphous with zircon is not substantiated. Outgrowths of secondary zircon on well-rounded grains lacking any remnants of crystal faces always have the same orientation relative

to the host material. These outgrowths of zircon on zircon are in most respects comparable to the examples described and figured by Grout and Thiel (Tyler *et al.*, 1940, Pl. 2, fig. 1; also p. 1491), but they do not reproduce the perfection of development of the forms described (Butterfield, 1936) from the Millstone Grit of Yorkshire, or (Smithson, 1937, 1941) for material from mid-Jurassic sediments of Yorkshire. These examples are, perhaps, more comparable still to the outgrowths on purple zircon recently described by Bond (1948, p. 36–38) from the Lower Wankie Sandstone of Southern Rhodesia.

A wide range of hyacinth crystals selected for determination of optical properties lay within narrow limits: O = 1.920 – 1.929; E = 1.967 – 1.981; E – O = 0.040 – 0.054. These values are distinctly higher than those recorded by Tyler and Marsden (Tyler *et al.*, 1940, p. 1441) for hyacinth, and partly covers the ranges of refractive index defined by Morgan and Auer (1941) for hyacinth and normal zircon.

Measurements of the optical constants for different zones of zoned hyacinth crystals were generally unsatisfactory, but for one crystal, which has a dark-purple central zone surrounded by a paler border and had been fractured normal to the *c* axis, refractive indices, for both sub-individuals, are as follows:

	Inner zone	*Margin*
O =	1.908	1.927
E =	1.930	1.976
E – O =	0.022	0.049

The refractive indices for both zones, except for that of the extraordinary ray of the outer zone, conform to the limits set by Morgan and Auer (1941, p. 308) for hyacinth, and, on account of the relatively low birefringence, the composition of the central portion of the crystal must be approaching that of malacon.

From mere inspection and evaluation by the Becke bright-line test, the central portions of the hyacinth crystals frequently appear to have lower refractive indices than the marginal zones. Actually this is contrary to the findings of Tyler and Marsden (Tyler *et al.*, 1940, p. 1442) since their experience showed hyacinth zoned externally by malacon but never malacon surrounded by hyacinth.

All of the inclusion-free hyacinth crystals tested have more or less normal densities that show a range of 4.60–4.62. Hyacinth crystals enclosed a variety of inclusions and, the number of inclusion-free crystals is generally greater than for colorless zircon. These inclusions are similar to those already found in colorless zircon except that xenotime, mullite (?), and cassiterite have not been recognized. Tantalum- and (or) niobium-bearing minerals were more often detected in hyacinth crystals, and particularly in those crystals exhibiting excessive micro-fissuring.

The general properties of the hyacinth found in New Zealand sands are quite comparable to those found by Tyler *et al.* (1940) in the Precambrian rocks of the Lake Superior region and the mineral described so clearly by Mackie (1925 a, b) that was derived initially from the Lewisian Gneiss of northern Scotland. To make quite certain that this comparison was valid, the heavy mineral fraction was separated from a brown Torridonian sandstone that the author had collected some years ago from an outcrop just south of Loch Maree in Ross-shire, Scotland, and shown by Mackie to contain plentiful hyacinth. In these hyacinths were found all the features of zoning, micro-fissuring, and rounding to be seen in the New Zealand material and distinction between them was impossible.

The difference in the degree of rounding of both normal zircon and hyacinth has already been commented upon and a method to give some measure to this characteristic without actual determination of roundness or sphericity was sought. It should be remembered at this point that the heavy mineral residues studied were obtained either by sporadic pannings or from dredge concentrates from river silts and gravels, the latter derived in the first place from extensive and heterogeneous distributive provinances and not from specifically located rocks. Therefore, methods involving a high degree of preciseness would provide data of little or no actual value and would take unwarranted time. The methods of both Trowbridge and Mortimore (1925) and Tickell (1947, p. 6) were used at first but later discarded in favor of a simplication of Allen's (1944, p. 72) scheme whereby a factor known as the zircon abrasion index is determined. The zircon abrasion index equals the percentage of euhedra among the zircons,

and to facilitate determination of this characteristic, zircon concentrates were prepared from each specimen by running the latter through the strongest magnetic field obtainable, after suitably sizing the material by screening and removing all ferromagnetic grains. Such concentrates often contained other practically nonmagnetic minerals such as apatite, cassiterite, scheelite, gold, uranium-free thorium silicate etc., but the percentages of these minerals were usually too small to interfere with the work. A portion of the material thus concentrated was mounted in hyrax and the grains counted according to the methods already detailed. Crystals exhibiting any sign of abrasion were not counted as euhedra, and in most instances at least 400 grains were counted; however, any preparation with less than 100 grains was not considered to give a significant figure. These Z_{a1} factors also clearly show (Table 18) that there is a distinct tendency for larger crystals to become rounded to a greater degree than smaller crystals of the same mineral. However, this can only be considered a generalization in this instance since there are a number of obvious divergences. In the zircons from Snowy River, an increase in the factor Z_{a1} with increase in grain-size is due primarily to an influx of larger colorless zircons of distinctive habit, a detail already commented upon. Similar circumstances may account for some of the anomalies found in other analyses. However, determination of the factor Z_{a1} can have only limited value when applied to minerals in concentrates obtained from Recent and heterogeneous river gravels since such specimens do not represent *en masse* results of careful sorting and fractionation. However, more detailed work is being carried out to evaluate the real usefulness of the abrasion index factor for material from Recent gravels such as Grey River and its tributaries.

TABLE 18.—ZIRCON ABRASION INDEX FOR TWO SIZE FRACTIONS

Sample	Normal zircon. < 62 microns	Hyazinth. < 62 microns	Normal zircon. 62–125 microns	Hyazinth. 62–125 microns
Slab Hut 1	25	9	18	8
Slab Hut 2	23	10	19	7
Grey River 1	18	16	14	9
Grey River 2	19	18	15	8
Grey River 3	20	17	14	9
Snowy River 1	17	11	22	10
Snowy River 2	19	11	23	9
Atarau	14	7	15	6
Ngahere 1	17	9	11	7
Ngahere 2	16	10	12	7
Gillespie's Beach	25	✱	15	✱
Okarito	26	✱	16	✱
Naseby	20	✱	✱	✱
Kyeburn	22	✱	19	✱
Lowburn Ferry	20	✱	17	✱
Ligar Bay	30	✱	25	✱
Kaniere River	13	✱	16	✱
Arahura River	15	✱	16	✱
Barrytown	14	✱	7	✱
Rimu	17	✱	15	✱
Red Jacks	17	✱	11	✱

✱ = Insufficient grains to enable an accurate count to be made.

ZOISITE: Zoisite has been found in a number of samples and is usually associated with the material of finer grain-size. It occurs as subhedral to anhedral grains that are either colorless or very pale yellow with distinct but weak pleochroism. Occasional semi-opaque grains with inclusions of dust-like particles of iron oxides, together with the somewhat ferriferous nature of the minerals themselves, made it possible to segregate them electromagnetically. A perfect cleavage, presumably (010), was noted (Pl. 4, fig. 59), and a majority of the crystals are tabular parallel to (100); as a result, there is a tendency for such crystals to orient themselves in mounting media so that the acute bisectrix is normal to the glass slide. A characteristic strong dispersion of the optic axes, with $\rho > v$, produces strongly anomalous interference tints when viewed between crossed nicols. A range of 33°–38° ± 2° was found for optical axial angles in ten grains, in which the optic axial plane was perpendicular to the cleavage; a positive optic sign was determined. Such properties agree closely with those listed by Winchell (1933, p. 311) for β-zoisite and are almost identical to those found for zoisite in heavy mineral assemblages of certain eastern Otago Tertiary and Cretaceous sediments (Hutton and Turner, 1936, p. 269).

In the finest size-fractions of some of Grey

TABLE 19.—MINERAL FREQUENCIES

| Grain-size (U. S. S. Screens) | Locality and Nature of Specimens | Minerals | | |

| | | |

River concentrates, a second distinct type of zoisite was observed. Morphologically similar to the variety just described, it exhibited distinctive optical properties. The grains were clear and colorless, there was strong dispersion of the optic axes with $\rho > v$, but 2V was too small to observe separation of the isogyres in white light. For $2V_{Na}$ an angle of $8° \pm 2°$ was obtained with the optic axial plane perpendicular to the cleavage (010); in blue light, however, the mineral appeared to be uniaxial. By virtue of the type of dispersion and position of the principal plane, this mineral is also β-zoisite with the tenor of iron less than that in the β-zoisite described above. In composition and optical properties, the zoisite with the small optic axial angle is almost at the transition point β-zoisite$\longrightarrow\alpha$-zoisite.

Nature and Cause of the Metamict State

The term metamict was introduced by W. C. Brögger to denote the phenomenon of isotropy or the glassy-amorphous condition sometimes found in zircon and in other normally anisotropic minerals. This state is now generally understood to involve in minerals of appropriate composition the production of a condition of complete structural isotropy in otherwise crystalline compounds; at the same time a general decrease in density and refractive indices usually becomes apparent, and a considerable degree of hydration is also to be expected. This physico-chemical transformation or metamictization takes place only in minerals containing radio-active elements but not necessarily in all of these; notable in the latter category are monazite, thortveitite[30], titanite, and possibly xenotime. A number of unsatisfactory hypotheses have been advanced to explain the mechanism of the process (Mügge, 1922; Goldschmidt, 1924; Vegard 1927), but Ellsworth (1925; 1932a) has put forward a more acceptable theory involving autoxidation, applicable particularly to uraninites but which may be extended to all minerals originally containing uranous oxide. According to this hypothesis, the disintegration of both uranium and thorium atoms gives rise, in addition to helium and appropriate atoms of the uranium and thorium disintegration series, to two oxygen atoms that immediately oxidize neighbouring uranous ions to the uranic state, and hence according to Ellsworth (1925, p. 138) "all minerals containing UO_2 automatically and irresistibly oxidize themselves at a rate depending on the rate of uranium and thorium disintegration". Tomkeieff (1946, p. 700–701), points out that if autoxidation were the only cause of a disrupted crystal lattice, a very clear correlation between the amount of lead and uranic oxide should be evident from analyses of uraninites. To evaluate this statement Tomkeieff has plotted PbO against the relative proportions of UO_2 and UO_3 for 52 analyses of uraninites; he found there is a wide scatter of points, but when the analyses are averaged into three groups on the basis of lead content a progressive increase of UO_3 with increase of PbO is apparent, and, the curve of these averages is approximately parallel to the curve that may be deduced theoretically. This would certainly appear to substantiate Ellsworth's hypothesis, so far as uranous-bearing minerals are concerned, and as Tomkeieff puts it "autoxidation may be a very real factor in transformation of uranous into uranic oxide". As contributing factors to the ultimate results of the autoxidation process, Ellsworth suggests that competition for oxygen between later members of the uranium and thorium disintegration series, and also the force of recoil of a radioactive atom ejecting an alpha-particle must inevitably tend to produce disruption of structure. The force of recoil is a factor of considerable magnitude. Since the alpha-particle has a mass of 4 and the uranium atom a mass of 238, the latter will have approximately a sixtieth of the energy of disintegration, probably in excess of that necessary to break the chemical bonding and allow the daughter atom to project itself some distance through the crystal lattice and there take up the configuration demanded by its chemical nature.

There is little doubt that the metamictization of radioactive minerals must be due ultimately to the presence of uranium and (or) thorium ions therein, and most illustrative in this connection is the work of Stackelberg and Rottenbach (1940, p. 207–8), who found that a thin plate of zircon when subjected to a radioactive

[30] One specimen was found to contain only 0.09 per cent ThO_2 (Marble and Glass, 1942, p. 697).

source for a period of 4 months broke down completely. And the data of Morgan and Auer (1941, p. 307) show that the birefringence of zircon is approximately inversely proportional to the measure of radioactivity for that mineral. However, the presence of radioactive elements in a mineral is not the sole condition for isotropization. If it were such minerals as xenotime, monazite, thorianite, and titanite should sometimes be found in a metamict condition. Thus some instability in the structure of the minerals may also be necessary. In his investigation of the metamict state Faessler (1942) considered that the transformation was probably due to alpha-radiation from radioactive ions, but that such transformations are more likely in minerals with complicated ion substitutions than in chemically simple compounds, which would probably be more stable. Although this may be the case for a number of radioactive minerals this is definitely not valid for zircon, chemically a relatively simple compound, while such minerals as monazite, xenotime, titanite, in which considerable substitution is generally evident, appear to be quite stable and are never, so far as the writer is aware, found in a metamict condition. Thus Faessler's supposition is not entirely acceptable.

Both Machatschki (1941, p. 38–39) and Tomkeieff (1946, p. 702) have offered hypotheses that may partially explain the phenomenon of metamictization in particular instances on the basis of structural instability. Machatschki considered particularly the structural relations in zircon and pointed out that the zirconium-oxygen distances are irregular, that is four oxygen atoms at 2.05 Å. and four at 2.41 Å. Let us examine the structural relationships in zircon, thorite, and xenotime in order to determine, if possible, whether structural instability is a necessary prerequisite for a change to a metamict state. The irregular distribution of the oxygens around the Zr^{4+} ions would appear to be an unstable arrangement, since the configuration of maximum stability would have the anions as close as possible to the central cation (Pauling, 1933, p. 1896); and theoretically the zirconium-oxygen distance should be approximately 2.27 Å. when corrected for eight-fold co-ordination and assuming a factor of 10 for the Born exponent since the former ion probably has krypton configuration. Further the radius ratio of the Zr^{4+} ion with O^{2-} is 0.62 using univalent ionic radii, a figure usually considered too low for eight-fold co-ordination. However, with this value, Zr^{4+} may be considered to occupy a transitional position since co-ordinations of six ($CaZrO_3$, zirconiferous pyroxenes), seven (ammonium zirconium heptafluoride), and eight (zircon and baddeleyite, the latter with a deformed fluorite structure) are all known. And since, in a completely metamict zircon, zirconium is found in the form of baddeleyite[21], a co-ordination number of eight may be the most stable configuration for that ion, and this arrangement, therefore, appears to be set up in zircon under the impetus of alpha-ray bombardment.

In the xenotime lattice, on the other hand, the observed inter-ionic distances between Y^{3+} and O^{2-} are very nearly identical, 2.23 Å. and 2.24 Å. (Vegard, 1927, p. 511) but nevertheless they are distinctly less than the corrected sum of the appropriate crystal radii, 2.41 Å. Naturally it is difficult to calculate the exact value of the equilibrium inter-ionic distances, since all the pertinent factors are not known, but because of the discrepancy between observed and calculated values, the Y^{3+}–O^{2-} bonding may be partly co-valent, in this instance at least. The ratio of the univalent radii is 0.68, not a transitional value (Pauling, 1927, p. 785) but one that would readily allow at least a square antiprism co-ordination. Therefore, on the basis of radius ratio and anion-cation distances alone, the structure of xenotime would appear to be more stable than that of zircon, and because of this stability apparently, xenotime does not become metamict.

Except for the thorites from southern Westland described earlier in this paper and one other obscure instance[22], the author has found

[21] The following compositions for completely metamict zircon have been observed by X-ray diffraction methods: ZrO_2 and tridymite, Borowsky and Blochin, 1936; microcrystalline cubic ZrO_2 and amorphous SiO_2, Chudoba, 1937a; amorphous ZrO_2 and amorphous SiO_2, Chudoba, 1937b; monoclinic ZrO_2 and amorphous SiO_2, Stott and Hilliard, 1946.

[22] Professor Adolf Pabst has pointed out to the writer that Boldyrev *et al.* (1938, p. 87–8) have published a thorite pattern and from it calculated the mineral's cell dimensions, *viz.* $a = 6.315$ Å; $c = 5.667$ Å. Obviously these workers have used anisotropic (= unmetamict) material for their X-ray study, but unfortunately they have not listed any other of its physical properties.

no record of that mineral in a completely unaltered condition; Brögger (1890, p. 124) referred to optically uniaxial, positive material, but this was in the form of minute relicts within highly altered thorite. Therefore, except for Boldyrev's material only metamict varieties of thorite were available for X-ray diffraction studies (Vegard[a], 1916, p. 93) and, accordingly, no structural data are available. However, from a theoretical standpoint, since $ThSiO_4$ appears to crystallize in the same way as zircon, one might reasonably assume that the former mineral consists of silicon and oxygen tetrahedrally arranged and Th^{4+} ions surrounded by eight O^{2-} ions; the Th^{4+}– O^{2-} equilibrium distance can only be assumed from the sizes of the appropriate ions. On the basis of Zachariasen's (1948, p. 1104) computations, the Th^{4+} ion has a radius of 0.95 Å., which differs unimportantly from that of Y^{3+}, whereas the radius ratio with oxygen of 0.67 should permit fairly stable antiprism configuration of the Th^{4+} ion.

Hence, there appears to be little structural reason why thorite should be so readily altered and a thorian-xenotime apparently stable; we must wait until the actual structural relationships have been determined.

Tomkeieff's (1946, p. 702) theory to explain the phenomenon of metamictization was applied specifically to the break-down of uraninite, that is for the change $U^{4+} \rightarrow U^{6+}$; he contended that this increase in the valency state was bound to decrease the ionic radius of the ion concerned. While such argument is correct in a general sense, the writer believes that Tomkeieff's assumption of the lowest possible radius ratio value of 0.225 for U^{6+} is unwarranted. There is no necessity to point out that within a range of 0.225–0.414 a tetrahedral configuration is the stable one, and in fact in some instances, the figure of 0.414 is not necessarily a maximum one; for example the Ge^{4+}– O^{2-} ratio is 0.43, yet the former element may be tetrahedrally co-ordinated as in $GeCl_4$ (Wells, 1947, p. 282). Of course the Ge^{4+} ion has a transitional value since GeO_2 is dimorphous occurring in both four-fold and six-fold co-ordinations, similar to quartz and rutile structures respectively. Therefore, Tomkeieff's figure of 0.30 Å. for the size of the U^{6+} ion is most doubtful, since the size of that ion might range anywhere from 0.31 Å. to 0.60 Å., but probably is closer to the higher figure, when the radius of the O^{2-} ion is taken to be 1.40 Å. and not the value 1.32 Å. used by Tomkeieff. Nevertheless, the volume change involved in the oxidation of $U^{4+} \rightarrow U^{6+}$ is considerable, that is, at least a change from 0.95 Å. to 0.60 Å., and from this point of view alone Tomkeieff's speculation would seem to be valid.

Finally, it is not clear in some cases whether the presence of uranium in a thorium-bearing mineral is a prerequisite for a radical transformation in the structure of an otherwise stable mineral. For instance the completely unaltered thorium silicate from southern Westland is uranium-free, but apparently this is not a common condition (Rutherford, 1906, p. 177); however, in this particular case the biaxiality of the mineral leads one to conclude that it is not tetragonal but has a lower symmetry that might be the stable one for thorium silicate. The presence or absence of uranium cannot possibly be the sole requirement since the uranothorite from southern Westland with 11.5 per cent of UO_2 shows only slight alteration and monazite may contain considerable uranium and yet is never found metamict. It is the writer's opinion that this can only mean that a monoclinic symmetry is the stable one for cerium phosphate; one is inclined to speculate on the possibility of the occurrence in nature of cerium phosphate with an unstable tetragonal symmetry. Alternatively are these southern Westland thorites not old enough for the transformation to a metamict state to have taken place in spite of an age determination of approximately 119×10^6 years for one of them? Certainly the metamict minerals of southern Norway and some Canadian areas are Precambrian, but this question will have to await an answer.

Origin of Micro-fissures in Radioactive Minerals

The curious arrangement or network of cracks, such as occurs both in uranothorite and hyacinth is by no means uncommon, and

[a] Vegard states that although several very fine crystals were tested not one of them gave any X-ray pattern at all. Thus the lattice must be completely broken down.

has been described by numerous investigators. However, such features appear to be restricted to radioactive minerals or at least have their origin therein, because from them the anastomosing systems of cracks radiate out into surrounding quartz, feldspar, mica, or other associated minerals. Landes (1932, p. 390) cites one instance where fissures radiate out 4 or 5 feet from the radioactive minerals concerned, and follows Hess and Wells (1930) in believing that the actual primary growth of the minerals, as a result of the increase in volume, shatters the enclosing but earlier crystallized minerals. In an early paper Ellsworth (1922, p. 58) stated

> "Another very characteristic indication of radioactive minerals is the presence of small cracks or fractures in the rock radiating from the mineral in all directions. These radiating fractures occur in feldspar, quartz, or mica surrounding radioactive minerals and are apparently caused by pressure which the mineral has exerted outwardly, due possibly to increase of volume resulting from internal radioactive changes."

Later, however, for a particular case he (Ellsworth, 1927a, p. 367) believed that the fracturing of radioactive minerals and enclosing matrix alike was the result of "the internal strains set up by autoxidation", but did not amplify this further. Walker and Parsons (1923b, p. 27–28) stated that shattering resulted from either (1) expansion of the inclusion, or (2) contraction of the host. They considered that expansion may result from oxidation and hydration, although they were careful to point out that such chemical changes did not always result in volume increase. On the other hand, where two minerals cooled from the same temperature and the coefficient of expansion of the host was greater than that for the included mineral, Walker and Parsons believed that the host would be very likely shattered while the inclusion might remain undamaged.

Without doubt the shattered or fractured condition of radioactive minerals and their environment results from expansion of the radioactive minerals concerned, but if growth or different physical properties are the cause, shattering should be found just as frequently about non-radioactive minerals; this does not appear to be the case. Therefore, in the writer's opinion, such fissuring may be attributed to expansion of the minerals concerned as a result of the transformation of radioactive minerals into the metamict state, whether complete or not. This physico-chemical change almost invariably appears to involve a notable decrease in density of the original minerals, and this is accompanied *pari passu* by considerable hydration and volume increase. Marked degradation of the crystal lattices also develops as a result of these changes as is clearly shown by Vegard (1916, p. 65) in the case of thorite and by Stott and Hilliard (1946), Chudoba (1937a, 1937b), and Bauer (1939) for zircon.

Acting singly or together upon anisotropic material, and the term anisotropic is used here in the widest sense, such changes could produce tensional and compressional forces competent to cause rupture of the minerals concerned; or if the radioactive material were present as inclusions within other minerals, the fissures would almost certainly extend outwards into the host minerals, or even beyond these into the surrounding matrix of adjoining crystals. Conditions of this nature could very well have produced the characteristics seen in the hyacinth crystals described, for in one example the central part of a hyacinth crystal clearly showed alteration towards a metamict state. Further, the detection of tantalum and (or) niobium in inclusions in zircon suggests the presence of minerals that are notoriously unstable and almost invariably found in a metamict condition.

Another condition frequently associated with radioactive minerals although naturally not observed in this study, is the red color, seen particularly in feldspars, of the minerals in close proximity to radioactive ones. Ellsworth (1922, p. 57) considered this feature an infallible guide to the presence of radioactive minerals but, although a pink coloration adjoining radioactive minerals is frequently found in pegmatites and similar rocks, the writer does not agree that it is always found. For example the orthites of the granites of Pomona Island (Turner, 1937a, p. 88–89) and in the pegmatites of Wilmot Pass (Turner, 1937b, p. 243) in Fiordland, the titanites in the north-west Nelson granites, and the monazite and xenotime in migmatites of the Charleston-Fox River area, do not appear to have produced a red color in the adjoining feldspars. It is perhaps significant that none of these minerals is altered to any extent, whereas the red coloration seems to be developed when

the uranium or thorium-bearing minerals are metamict or in a condition approaching that state. Therefore, it is suggested that the red coloration may be developed in the following manner: The transformation of a radioactive mineral to the metamict state will result in fracturing of the mineral and the micro-fissures may in some instances extend into the surrounding matrix. These fractures will provide a ready means for solutions to circulate and facilitate further break-down of the metamict mineral with subsequent removal or dispersion of decomposition products. Most uranium- or thorium-bearing minerals contain iron, and the oxidation of this constituent in particular would produce a zone of coloration surrounding the radioactive mineral. It may be significant that distinctively colored zones surrounding radioactive minerals appear to be developed in pegmatites etc. that are geologically old; that is, sufficient time has elapsed for transformation to a metamict state to be brought about.

Distributive Provinances

The investigation of over 400 samples of beach and dune sands, artificial and natural concentrates from alluvial and eluvial sands and gravels, has enabled the writer to obtain a generalized picture of the nature and distribution of heavy minerals in several areas of the South Island, and particularly in Westland. It is freely admitted that this study is incomplete, but it should point out some of the problems pertaining to the nature and distribution of heavy detrital minerals in a region of very great interest and the highest complexity. The actual source of many of the minerals described is unknown in some instances, owing chiefly to the absence of detailed petrographic and mineralogical studies in mountainous and often almost inaccessible country. In spite of this, however, some of the facts will at least assist further prospecting for rare earth and other valuable minerals, and provide information as to the nature of the minor accessory constituents in a wide range of local rocks that may assist in general stratigraphic problems.

In a very general sense, Westland and southwestern Nelson Provincial areas can be divided into three distributive provinces in which distinctive detrital mineral assemblages occur. These are:

(1). The southern to central Westland area extending from Bruce Bay northwards to about Saltwater Creek and Harahari in which a zircon-scheelite-thorite-uranothorite assemblage is characteristic.

(2). The Ross-Lake Mahinapua-Kokatahi-Arahura area with extensions north towards near Kumara, characterized by a relatively simple zircon-garnet-epidote (clinozoisite)-hornblende assemblage.

(3). The Grey River basin and the Cape Foulwind-Westport coastal plain in southern Nelson which is characterised principally by a zircon-hyacinth-monazite-xenotime-tantalian cassiterite assemblage.

Although the mineral assemblages of southern to central Westland are relatively simple they contain three unusual minerals, scheelite and two varieties of thorium silicate, which taken together with the tourmaline, topaz, and kyanite[34] present seem to have been derived in the first instance from the belt of granitic and metamorphic rocks that make up the western flanks of the Alpine Chain and occur as inliers on the morainic-covered coastal area. The abundant garnet is spessartite-almandine of average refractive index 1.791, and is quite comparable optically and chemically with the garnets so common in pegmatitic dykes and sills that outcrop here and there from the Mahitahi and Paringa rivers southwards. The pegmatites in this vicinity are generally the simple quartz-feldspar-mica type and quite distinct from those that normally carry rare earth and radioactive minerals, or other pneumatolytic-hydrothermal phases.

Uranothorite occurs largely as euhedra of approximately the same order of grain-size as the associated zircon crystals, which definitely suggests that in its original habitat the mineral was widely disseminated, and its occurrence today is not the result of the break-down of economically important segregations or concentrations of large crystals in the source rocks. On the other hand, the uranium-free biaxial thorium silicate invariably in the form of fragments and quite devoid of original crystal faces, must have originated by the break-down of crystals decidedly larger than the fragments

[34] Boulders of muscovite-kyanite schists are not uncommon in the alluvium of Paringa River.

now found, but this does not necessarily indicate that economically valuable segregations of thorite may exist at the source; in fact, the complete absence of even occasional fragments of thorite among the coarse material indicates that any large and valuable concentrations of this mineral are unlikely to be found.

Scheelite which occurs in payable veins in the Otago schists, principally at Glenorchy and Macraes, situated some 140 miles southeast and 160 miles south from the central Westland area, is a common detrital mineral in the sands of Central Otago streams. An extension of these Otago-type schists is to be found along the western side of the Southern Alps, immediately east of the central Westland area, and therefore, the detrital scheelite found here very probably has its source in such rocks.

Black titanite was found in a beach concentrate from Bruce Bay in the central to southern Westland area. In this sample, composite fragments coarser than 18-mesh consisted chiefly of fine-grained carbonaceous schist that almost certainly belongs to the Chlorite Zone. When isolated, crushed, and centrifuged in bromoform these schist fragments gave a heavy residue of "schist"-type tourmaline (Hutton and Turner, 1936, p. 264, 268; Hutton, 1940, p. 36, 63) and euhedral titanite crystals; the latter match those found in the finer size-fractions of the beach concentrate. There seems little doubt that this variety of titanite has come from either the Chlorite Zone schists that outcrop along the western flanks of the Southern Alps to the east and south of Bruce Bay, or the schistose Greenland Series that outcrop immediately south of this locality.

Dunites, harzburgites, and other olivine-rich rocks outcrop in a narrow, 40-mile long belt on the western slopes of the Olivine Range, some 60–70 miles southwest of Bruce Bay. Most of the detritus from these rocks reaches the coast by way of the Gorge, Cascade, and Jackson rivers, and the olivine in an isolated sample from Bruce Bay probably comes from these sources, possibly added to by a very small but indefinite amount through contributions from basaltic and other basic volcanic material associated with mid-Tertiary marine sediments (Henderson, 1917, p. 106) that outcrop along the coast southwards from the mouth of the Paringa River. Although both chromite and awaruite are found in the peridotites of the Olivine Range neither of these minerals was recognized in the mineral assemblage of this olivine-bearing concentrate.

The central Westland mineral assemblage has been found in numerous localities in the wide belt of fluvio-glacial sands and gravels that make up the sloping ridges and coastal plain fronting the main range. This alluvium has been re-worked time and again by the sea, and the heavy mineral assemblages concentrated therefrom have been buried to form fossil placers as the coast line advanced seawards. At the present day, particularly under storm conditions, the heavy mineral assemblages in the coastal dunes may be further concentrated to produce pay streaks and black sands on the beaches that are, at irregular intervals, valuable sources of gold.

The second area, characterized by the distinctive mineral assemblage zircon-garnet-epidote (clinozoisite)-hornblende with marked predominance of the last two minerals in a number of instances, suggests a metamorphic source. Undoubtedly however, the Tuhua granites have contributed a good deal of the hornblende, epidote, and clinozoisite since these intrusives have been much affected by cataclasis with the development of considerable epidotization of the plagioclase, alteration of amphiboles to chlorite, epidote, and zoisite, and production of semi-opaque titanite from opaque iron ores; the granite outcropping near the headwaters of the South Branch of Humbug Creek (1¼ miles N.N.E. of Frazer Peak) in Totara Survey District, Mikonui Subdivision, is typical in this respect. The occasional crystals of uranothorite, topaz, and cassiterite present in the sands and gravels of this area may possibly have the same source as the epidote, zoisite, hornblende, etc., but more probably have been transported northwards from the Bruce Bay-Gillespie's Beach area by along-shore currents and incorporated with the more locally derived material during deposition of the fluvio-glacial debris that makes up the coastal plain. Although biotite is very poorly represented in the samples studied, it is not as rare a constituent as it would appear since both muscovite and biotite are generally abundant throughout the sands and gravels of this area. The low density and flaky

nature of biotite tend to prevent it from being concentrated either on the gold-saving tables of dredges or in the gold pan. The paucity of tourmaline might appear somewhat unusual since some of the detrital material has been derived from granites. However, work on the Tuhua granites shows that this mineral is not commonly developed; notable exceptions in this respect are outcrops at the northern end of the intrusion in the vicinity of Te Kinga Peak and Hohonu Ridge, just south of Lake Brunner, and isolated and strictly local occurrences in Totara Survey District, Mikonui Subdivision.

The Grey River-Mawheraiti River basin and the triangular coastal plain lying south and east of Cape Foulwind are characterized by a third and distinctive assemblage of zircon-hyacinth-monazite-xenotime-tantalian cassiterite; concentrates from Atarau dredge, although they distinctly show the general characteristics of the Grey River area, are heavily contaminated with material from a different source than that which provided the bulk of the gravels and sands of the Grey River and Cape Foulwind areas. However, tracing the various detrital minerals to their real source is exceedingly difficult owing to an absence of any detailed knowledge of the mineralogy of the granites, schists, and gneisses that make up the main masses of the Paparoa Range west of the Grey and Mawheraiti Rivers, and the Victoria Range to the east thereof.

The dominant heavy detrital minerals in the sediments of the two areas have almost certainly been derived initially from granitic and gneissic rocks and the associated pneumatolytic-hydrothermal phases; these sources are most probably the rocks of the Paparoa Range and, before extensive capture of the head-waters of the important eastern tributaries of the Grey and Mawheraiti Rivers, the intrusives of the Victoria Range lying to the east. The distribution of these index minerals is by no means limited entirely to the Quaternary gravels of the two areas in northern Westland and southern Nelson but have been found in a wide range of rocks[35]. Clearly therefore, the source rocks have been providing a supply of such minerals for a considerable period. In an attempt to determine the actual source rocks of these minerals a wide, but far from complete, range of granites, gneisses, and schists from this region were examined, but monazite, xenotime, and cassiterite were not found in the schists and granites, except in rare cases when occasional crystals of the first two minerals were represented. A few rare grains of dusky brownish zircons also found in these rocks differed from hyacinth proper in color and crystal habit. However, the gneisses, particularly those outcropping along the coast south of Charleston and the mouth of the Nile River and those occurring as boulders in the upper reaches of the Otututu River, were of considerable interest. The minor accessory minerals from four gneisses collected at intervals along the coast south of Charleston averaged 4.7 per cent by weight, and the following percentages of the different constituents were found: zircon 33, apatite 37, monazite 26, xenotime 3, opaques 1; no hyacinth or cassiterite were present, however, in any of these assemblages. In similar gneisses from other outcrops in these two general localities, occasional grains of garnet, epidote, zoisite, tourmaline, and topaz were observed in addition to the characteristic minerals mentioned. Almost certainly these or similar gneisses, which are widely distributed throughout southern Nelson, must have been the main source of supply of the monazite and xenotime found in the Quaternary gravels, although undoubtedly these minerals have been incorporated in other rocks younger than the gneisses (the Trias-Jura rocks of the Hurunui River to mention only one instance) and the decomposition and degradation of these must also in their turn have provided some portion of the material supplied to the Recent and sub-Recent gravels. No indication of the initial source of cassiterite or the origin of hyacinth has been found, but in all probability, cassiterite came from the neighbouring granites and associated pegmatites. In this connection the hypotheses of Borovick and Gotman (1937, p. 352–353) and Boldyreva (1939, p. 421) that cassiterite from pegmatite veins contains enhanced quantities of tanta-

[35] The distinctive heavy mineral assemblage zircon-hyacinth-monazite-xenotime-cassiterite is the dominant one in conglomeratic rocks of Trias-Jura age outcropping in the Hurunui River area, Mandamus Survey District, about 75 miles E.S.E. of Grey River; also in Tertiary cementstones and conglomerates in Lankey Gully, Reefton Survey District.

lum[36] and niobium, suggest that the pegmatites rather than the massive granites or their more attenuated phases may be primary source rocks for Westland cassiterite.

It was pointed out earlier that hyacinth crystals exhibited a far greater degree of rounding than the associated colorless zircon; in fact grains of hyacinth that are almost spherical are not uncommon. At first sight the rounding of hyacinth would be attributed to abrasion, many times repeated, during a long and complex history since the liberation of the mineral grains from their ultimate source, but, in this connection, it should be remembered that other workers have noted that zircon, and particularly hyacinth, are often distinctly rounded and sometimes ovoid, although still in the host rock, and they need not show sharply euhedral forms (Armstrong, 1922, p. 395; Frankel, 1943, p. 77–78; Warner, 1945, p. 95). In this work crystals of hyacinth have not so far been found in their original source rocks, although some of the dusky-brown zircons in some samples of Paparoa granite did not show the perfection of form exhibited by associated colorless zircons.

In other localities, the original source rocks of hyacinth appear to be Precambrian (Scotland, Mackie, 1925b; North America, Bruce and Jewitt, 1936; Report on Accessory Minerals in Crystalline Rocks, 1935; Tyler *et al.*, 1940; Australia, Carroll, 1939, 1941, 1944; Higgens and Carroll, 1940; Madagascar, Lacroix, 1922). On this basis it is quite unsafe to suggest an initial source of Precambrian age for the hyacinth found in these southern Nelson and Westland sediments, if only because Precambrian rocks have not yet been recognized in New Zealand. Nevertheless, this possibility must be borne in mind.

Another mineral of some consequence occurring in the sands and gravels of this area is aluminian chromite, and it reaches its greatest abundance in the coarser fractions of concentrates from Blackball and Red Jacks dredges. The presence of chromite in this general area was noted by Morgan (1911, p. 91) who commented upon its relative importance in concentrates from Montgomery's Terrace Sluicing Claim. The immediate source of this mineral is not known, but outcrops of peridotites, serpentinites, and their metamorphosed equivalents, antigorite and talc schists, are not uncommon along the western slopes of the Southern Alps, and doubtless the mineral has been brought down the westward-flowing tributaries of the Grey and Mawheraiti Rivers. The relative abundance of aluminian chromite in samples from the immediate vicinity of Ngahere suggests that the Pohaturoha (or Big Grey) River, now beheaded by the Inanguahua River, may have been the vehicle for much of this material.

The mineral assemblages in samples from Atarau dredge are somewhat distinct from the type usual in the Grey River basin in that epidote and clinozoisite are generally as plentiful as zircon. This suggests that the immediate source of a considerable amount of the sands and gravels treated by this dredge is the Aorere greywackes, argillites, and schists of the southern end of the Paparoa Range.

The mineral assemblages in a number of concentrates from Barrytown dredge are typical of most of the paystreaks obtained from numerous Westland beaches from north to south. Barrytown material in particular contains such minerals as scheelite and uranothorite, and thus shows the influence of the northward-flowing off-shore current which has brought along the coast a small percentage of those constituents characteristic of the central Westland provinance, whereas the occurrence of monazite, xenotime, and tantalian cassiterite indicates a mingling of material typical of the northern Westland-southern Nelson region.

The paystreaks and black sands that have accumulated on beaches between Wainui Inlet and Separation Point in northwestern Nelson, of which the sample from Wainui Inlet is typical, have been derived from the Pikikiruna Range-Separation Point granites, that outcrop along these shores; the heavy detritals are especially interesting on account of the high percentage of titanite therein. Actually, the quantity of titanite in these granites is in excess of that found in normal granitic rocks, and the prominence of this mineral suggests contamination. Field examina-

[36] The possible occurrence of tantalite in Blackball Dredge concentrates (p. 670 of this work) and sluicing claim concentrates from Addison's, Westport (Dom. Lab. 50th. Rept., 1917, p. 25) should be noted in this connection.

tion of the relationships of the granites and closely associated marbles and quartzites leaves no doubt that much of these sediments has been metasomatized and otherwise assimilated by the granites at the time of their emplacement to produce a range of amphibolites and grossularite-bearing rocks of basic appearance. In many of these titanite is a prominent constituent. Even in those granites that appear to be removed some distance from obviously metasomatised derivatives, titanite may be clearly seen in hand specimens. Apatite is also plentifully represented among the detrital minerals concentrated on the north-west Nelson beaches; this feature may be correlated with the relative abundance of apatite in some of the metasomatized and contact-altered marbles.[37] Tourmaline is uncommon in Wainui Inlet and neighbouring sands in spite of the fact that the mineral is not rare in the Pikikiruna Range intrusives. A blue tourmaline[38], a member of the dravite-schorlite series, that occurs at Page's Saddle is typical, although clear-yellow elbaitic varieties have been found in these granites.

The mineral assemblages obtained from the alluvial sands and gravels of Central Otago are those to be expected in material derived from a region of low-grade dynamothermally metamorphosed graywackes, except for, in the Naseby-St. Bathans-Kyeburn area in particular, the quantity and nature of the detrital zircon. In the samples studied, zircon is clear, colorless, and almost invariably sharply euhedral, and since the schists are generally the metamorphosed derivatives of graywackes and similar quartzo-feldspathic sandstones, one might expect some degree of rounding in the zircons, or at least 5 or 10 per cent of fractured crystals. Such characteristics are only found in isolated crystals and cannot be considered more than a rare occurrence. Added to this is the relative abundance of zircon in the schists in the vicinity of Naseby and St. Bathans and, consequently, in the auriferous placer deposits derived therefrom. Zircon has usually been considered a very insoluble, stable mineral, and not of that group of minerals, chiefly silicates, that may be carried in solutions and ichors forming the vanguard of a regional intrusion of granitic material. The work of Goldschmidt (1919, p. 280–1), von Eckermann (1922, p. 375), Gillson (1925, p. 190), and Franco and Loewenstein (1948, p. 150) have shown beyond much doubt that zircon or zirconiferous material may be transported pneumatolytically. Concentration of zircon within limestone, but outside actual granite contacts in the Mansjö Mountain area described by von Eckermann, can have no other explanation. Therefore, to account for the plentiful, yet sharply euhedral zircon in localized areas of the Otago schists, the writer suggests that at least some of it has been introduced into the schists and, together with tourmaline (Hutton, 1939, 1940), represents one phase of emplacement of subjacent acid intrusives that was part of the general conditions culminating in dynamothermal metamorphism of the rocks of the Otago area.

Although granitic rocks do not come to the surface anywhere in the Otago Central area, there can be little doubt of their presence at lower levels in view of the wide and often plentiful distribution of tourmaline throughout the schists. And, further, the mineral assemblages in Tertiary sediments of Eastern Otago can only have been derived from rocks of the Otago schist type that have been locally metasomatized by granitic intrusions; this association of rocks is clearly shown by the occurrence in the heavy mineral assemblages prepared from some of these sediments of such minerals as kyanite and andalusite (Hutton and Turner, 1936) and corundum and spinel (in auriferous gravels at Adam's Flat, near Milton).

[37] Bands and lenses of metasomatized marble are well developed in many places along the Pikikiruna Range, and in a typical exposure at Rameka Creek, on the western slopes of the range, approximately 10 miles south-west from Separation Point, crystals of apatite up to 5 mm. in diameter are not uncommon. Similarly near the headwaters of Brooklyn Stream, near Riwaka on the eastern side of the range, apatite is also found. A careful study of such rocks might reveal apatite-bearing material of economic value.

[38] E = 1.632 ± 0.002; O = 1.659; O − E = 0.027; dichroism: deep indigo blue (O) to colorless (E).

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School of Mineral Sciences, Stanford University, California.

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TECTONIC SCARPS AND FAULT VALLEYS

By C. A. Cotton

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Abstract

Faulted and strongly deformed land surfaces are everywhere subject to rapid dissection; but bahadas or aprons of alluvium are not as commonly developed in association with tectonic scarps in humid regions as they are in the semiarid North American Great Basin. Undercutting by rivers leads on the other hand in some cases to considerable erosional modification of tectonic forms, so that the fault scarps of humid regions may come to bear but little resemblance to type forms as generally described and figured in textbooks.

The Awatere fault-scarp type, as found in New Zealand, is contrasted with the Basin Range type. Scarps of the Awatere type are not fringed by bahadas, and in some cases they have been undercut by rivers.

The category of tectonic scarps includes monoclinal as well as fault scarps and also combinations of monoclinal with fault features. Many fault-line scarps also are tectonic, but only such of these as have been exposed by erosion in a first postfaulting cycle.

The facets of a dissected tectonic scarp, though they may be very flat and strikingly coplanar, rarely preserve actual fault surfaces. They are landforms of normal development, usually graded, with a straight base line.

Fault valleys, as distinguished from purely erosional fault-line valleys, are common and are occupied by postfaulting consequent rivers. Asymmetric furrows formed initially between inclined block surfaces and the scarps of a fault-block complex may guide drainage, and the valleys in such major fault angles are in many cases long-lived. Relatively minor fault-angle valleys are formed in districts with considerable prefaulting relief only when fault scarps develop along the lines of pre-existing rivers; this is not unusual if some rivers are already fault-guided, flowing perhaps in fault angles of earlier formation or in fault-line erosional valleys.

Other consequent valleys may have been formed concurrently with a long series of intermittent lateral (transcurrent) movements, and besides these there must be some which are walled on both sides by fault scarps related to a single fault or fault zone. Where reversal of movement has taken place in successive faultings, opposed scarps may be present. Reversal of the dip-slip movement on steeply inclined fault planes is not uncommon among ancient faults, and reversal in very recent times has taken place on active faults in New Zealand. If movement in the new sense continues, trenches with the dimensions of valleys will be formed, potential fault valleys of an opposed-scarp type.

Introduction

Fault Scarps and Related Forms

Minor and even major landforms of strictly tectonic origin, common as they are in some regions, are not always easily recognizable. Discussion of criteria for separation of fault scarps (tectonic) from non-tectonic forms of similar appearance developed by selective erosion has languished since Blackwelder (1928, p. 292) pointed out that the former are very rare indeed throughout very extensive regions. Though he figures and describes fault scarps in California, Blackwelder reaches the conclusion that tectonic scarps are not as common even there as has been supposed.

One cannot, however, relegate all scarps which are difficult to diagnose to the non-

tectonic, erosional category—certainly not all those in the mobile belts. In some stable regions, on the other hand, competent authorities express disbelief in the existence of fault scarps (Johnson, 1929, p. 357; Baulig, 1935) or any other forms of tectonic relief. In "countries bordering the North Atlantic", we are told, "there is no local ridge or valley which owes its existence to concurrent deformation" (Willis and Willis, 1929, p. 312). Yet, even in France, Meynier (1940, 1947) has pointed out some well-worn fault scarps, and in the valley of the Meuse Macar (1939, p. 1023; 1946a; 1946b, p. 263) has shown that faulting has dislocated the surface to a small extent.

Contrasting with the more stable regions, "the Pacific borders are . . . alive. From Cape Horn via Alaska and Japan to the East Indies the coastal ranges are in active movement" (Willis and Willis, 1929, p. 312). Imamura (1939), though he rejects some reports of extensive block structure throughout Japan, finds evidence of strong surface buckling which can be interpreted only as an effect of active folding in compression—a current orogeny. New Zealand is another region of diversified relief much of which is tectonic. Even in Australia, where severe earthquakes are unknown and very weak seismicity survives only in certain districts, some major landforms are regarded as tectonic even by the most critical students (Cotton, 1949b).

It is obviously impossible to lay down hard and fast rules for distinguishing between tectonic and wholly erosional scarps, but the separation can often be made without much difficulty when all relevant facts are taken into consideration. Some knowledge of the structure and of local geological history makes a decision possible in many cases, and the recognition of such fault scarps as can be identified then throws further light on geological history. "The distinction between fault scarps and fault-line scarps . . . is not simply an academic question of systematic nomenclature" (Davis, 1913a, p. 206); as Johnson (1929, p. 356) has remarked, "it is the indispensable prerequisite to a correct interpretation of geological history."

Many examples are known of scarps—high scarps as well as the miniature features which Davis and Blackwelder (1928, p. 304) term "scarplets"—that are still infantile and have patently been formed at the outcrops of faults either still active or very recently moving. A low scarp of very recent origin has been reported even in the relatively stable Australian region; it has been formed so rapidly, apparently, that it has diverted the Murray River (Harris, 1938, pp. 45–60). Infantile scarps as much as 1500 feet high are confidently interpreted as the outcrops of faults intersecting active Hawaiian volcanoes (Stearns and Macdonald, 1946, p. 40). These scarps are either quite fresh and nearly vertical walls of hard lava rocks, or, when slightly worn back and furrowed by erosion, continue to resemble structural escarpments of hard horizontal or slightly dipping strata, with limited accumulations of rock debris forming talus slopes and alluvial cones along the base. A more advanced condition of basaltic erosion on fault outcrops is seen on scarps of the Hawaiian islands Lanai and Molokai (Stearns, 1940, p. 12, 1947, p. 31). The less worn of the Basin Range scarps in the basalt-plateau regions of southern Oregon (Russell, 1884, p. 436; Johnson, 1918, p. 229–236; 1929, p. 366; Fuller, 1931, p. 23–36) and western Idaho (Anderson, 1941, p. 215) resemble these, but are, on the whole, somewhat more maturely dissected.

Scarps in a still more advanced condition of dissection, though none the less fault *scarps* according to strict definition of the term, are more difficult to diagnose, and this is true especially of those developed by dislocation of terrains of homogeneous rock, such as granite, and of rocks of heterogeneous structure, including ancient welts of folded and deformed strata. Davis (1901, 1903) has emphasized the absence of any necessary relation between the structure of a folded terrain and those scarp features intersecting it that are consequent on faulting.

Criteria of landscape faulting in seismic regions have still to be developed and systematized. As yet merely a sketch can be presented, based mainly on personal observations in New Zealand, where all the major landforms, except accumulation forms such as deltas and volcanoes, are tectonic (Cotton, 1916b; 1918; 1922, p. 178–188; 1925). Minor features also

in parts of this region are clearly of tectonic origin, being attributable to recent and even current orogeny expressing itself in some places as faulting and monoclinal flexure and in others as warping so pronounced that it may be described as producing a "folded erosion surface" (Willis and Willis, 1929, p. 312). The theoretical objection raised by B. Willis (1938, p. 808) to such "loose" use of the term "fold" to describe "flexure of the surface of the earth" does not seem worth while maintaining.

In the southern part of the South Island of New Zealand landforms resulting from a late orogeny are very stable and long-enduring (Cotton, 1948b, p. 306); in the geomorphic districts of both islands adjacent to Cook Strait (Cotton, 1945, p. 46; 1946, p. 232–235), they are newer and, in the most seismic parts, still developing. Here, however, owing to the structure of the terrain and partly also to the tectonic pattern developed, the forms produced tend to fade out or be blurred by erosion more rapidly than in the south.

In these landscapes fault scarps are numerous, as is the case also in other large parts of the mobile belts of the earth, but many escape identification as tectonic forms, remaining either unsuspected or in doubt because fault scarps departing from a standard type have been little studied. The selection of types for comparison has not been carried very far, and most criteria available for recognition of tectonic forms in actively writhing and seismic belts have been borrowed from the North American Great Basin. The mention of fault scarps is thus liable to invoke a possibly misleading mental picture of a landscape made up of ranges and basins closely resembling those of Utah or southern Oregon. In the circum-Pacific region, however, not only are the orogenic mechanism and tectonic pattern it produces different, but scarps may pass through their erosional sequential development under contrastingly humid climatic conditions.

Diagnostic Characteristics of Fault Scarps

Distinction from Fault-line Scarps

Blackwelder (1928, p. 294) has pointed out that fault scarps and fault-line scarps tend by convergence of development to resemble each other very closely; he almost despairs of distinguishing between them morphologically. The features common to scarps of both kinds, though not necessarily all conspicuous in every example, have been listed by Blackwelder (1928, p. 294–299), but he assumes that an observer will base his judgment partly also on the whole aspect of the surrounding landscape and what is known of the structure of the terrain.

In many cases the topographic features, considered alone, afford strong presumptive evidence of the *presence of faults*, though they leave the question of *landscape faulting* undecided.

The paucity of conclusive evidence one way or the other is indeed all that Blackwelder insists on. "By stripping a weak formation from an underlying resistant one in a faulted sequence", he points out (1928, p. 299) "the processes of erosion may exhume a surface that simulates very closely the form produced by crustal dislocation. Under suitable conditions this eroded surface exhibits splintered scarps, steps, facets, V-shaped canyons, a linear base, truncated folds, and hanging valleys, as well as other features that are also induced directly by faulting."

In the list given by Blackwelder (1928, p. 299–305) of features which may be relied on more particularly as indicators of tectonic origin, or landscape faulting, some are to be found only where faulting is still in progress (those he numbers 1 and 8–11), but the fact must not be overlooked that there are many pronounced tectonic scarps aligned on faults which show no signs of continuing or even very recent activity. Item 7, "displacement of an older topographic surface", is a decisive criterion, but it can rarely be applied (R. Willis, 1925, p. 664). Blackwelder (1928, p. 305) remarks that some of the criteria listed are "seldom available" and that others are untrustworthy except in special circumstances. None of them affords complete solution of the problem. He recommends elimination of fault-line scarps from the doubtful list by application of six tests devised for their recognition (Blackwelder, 1928, pp. 305–308). Item 4 in this list, "proof that the fault movement was pre-Pleis-

tocene" calls for rejection, because it almost amounts to another assertion that there cannot be a fault scarp without active faulting on its line.

The fifth criterion, "little or no alluvial deposit on the downthrown block" is untrustworthy in regions with abundant rainfall. Many rivers flow in consequent valleys along the bases of fault scarps, and such conditions are not conducive to accumulation of bahadas, at any rate after the landscape has become mature.

Significance of Antecedent River Courses Across Fault Blocks

One of the best tests indicating erosional development of scarps is superposed drainage across scarp-bounded ranges (Blackwelder, 1928, p. 308). If similarly placed antecedent courses can be identified, they afford a criterion of tectonic origin of the ranges they traverse, and commonly of their bounding scarps also (Davis, 1924, p. 321; 1930, fig. 3, p. 22).

As examples of antecedent rivers thus surviving: "Witness the course of Sevier River directly through, instead of around, the northern part of the Canyon Range in southwestern Utah; witness also some remarkable instances of antecedent drainage in the tilted lava blocks of arid northeastern California" (Davis, 1930, p. 22; *compare* 1934, fig. 8, p. 190).

In New Zealand, an antecedent gorge followed by the Waiau River through an actively tilting block (Cotton, 1947c, p. 12) affords collateral evidence of recent tectonic activity along a scarp which will be described later. The same river crosses several other recently upheaved tectonic blocks in gorges in its lower course (Cotton, 1948b, fig. 290A, p. 383).

In southern Oregon, Johnson (1918, p. 233) has described a wind gap which notches the crestline of a fresh fault scarp; similar to this is Geary's Gap, notching the long and wall-like Cullarin scarp, at Lake George, in eastern Australia (Taylor, 1907, p. 327; 1910, p. 7).

Well-defined wind gaps across tilted-block ranges in the northern Rocky Mountains, and interrupted valleys traced and matched from block to block (Anderson, 1947, p. 70–73), strongly support the diagnosis of the ranges as tectonic in origin, if, as seems to be the case, the gaps have been cut by rivers defeated only after persisting as antecedents during early stages of upheaval. If the gaps were, on the other hand, relics of an incomplete prefaulting dissection due to rejuvenation of rivers then in existence, they would afford equally good proof of dislocation of the land surface (Blackwelder's test, "displacement of an older topographic surface").

One or other of these explanations if applied to the Pass Creek air gap, Utah, described by Gregory (1949, fig. 3, p. 980, 996), confirms upheaval of a block bounded by Clark Mountain fault.

An upthrust horst forming the Stirling Ranges, in the south of Western Australia, which Woolnough (1920, p. 110) has described as bounded on the north and south by high fault scarps, is cut into segments by several wind gaps so deep that their floors are but little higher than the surrounding undisturbed (or little disturbed) peneplain. These afford passes which "traverse the range completely and afford very easy passage across the mountain area" (Woolnough, 1920, p. 83). It has been suggested that antecedent rivers persisted long enough to cut these passes (Jutson, 1934, p. 89, 272).

Each such case must be examined in relation to the geomorphic environment, and an alternative hypothesis of dismemberment of superposed river courses in the cycle now current must be considered. In a discussion of transverse courses followed by a number of parallel streams through a faulted block forming the "front hills" of the Haldon Hills, New Zealand, a hypothesis invoking superposition of a group of consequent courses has been preferred to that of antecedent origin (Cotton, 1938, p. 195–198).

Fault Scarps and Their Sequential Development

Development in a Semiarid Climate

Fault-made landscape forms developed in relation to great normal faults in the North American Great Basin have been subject to erosional gradation in a special climatic environment. The erosional power and transport-

ing capacity of rivers in the lowlands between ranges being comparatively weak under semi-arid conditions, abundant debris derived from dissection of a scarp-bounded range accumulates in the foreground of the scarp, in front of its base, to form a piedmont alluvial fringe, or bahada. This fends off from the foot of the slope any consequent river that might under other conditions flow longitudinally there in the asymmetric furrow, conveniently termed a fault angle, between a scarp and the inclined surface of an adjacent block. A river thus pushed back from the fault line or from the tectonic axis of a basin, fault angle, or asymmetric synclinal depression will, when later it cuts a valley, thereby fix itself permanently in a course perhaps far removed from this axis and the fault line (Davis, 1913a, p. 195).

Fault Scarps of the Basin Range Type

As a result of early discovery and recognition in a relatively waterless region, the fault scarp with such a specialized history became the type, especially after block diagrams of Basin Range scarps had been drawn by Davis, to be copied later by textbook writers. Fault scarps of this Basin Range type, as it may be called, are recognizable in some other parts of the world; but it is necessary to examine the development of tectonic forms in humid regions as well, where, primarily because of the vastly greater transporting capacity of rivers, erosional modification of scarps both during and after tectonic growth proceeds on different lines, ultimately producing a major landform with only a family resemblance to Basin Range fault scarps.

According to Davis (1913a, p. 197):

> It is evident that as long as waste is abundantly washed down from the uplifted mass no continuous stream and valley are to be expected along the fault line, even in the more advanced stages of the new cycle, unless highly specialized initial conditions have favored their development; and as such specialized conditions must be rare, the occurrence of consequent streams and valleys along fault lines in a youthful stage of the cycle introduced by faulting must be rare.

Later, when defining a scarp-making Wasatch type of fault (distinguished from a non-scarp-making San Andreas type, Davis (1927, p. 669) described the profile produced characteristically by dip-slip, or "Wasatch", faulting as follows:

> The side of elevation is occupied by a more or less dissected mountain range, while the side of depression is occupied by an aggraded plain of mountain-supplied detritus.

The conventional fault scarp.—The conventional fault scarp of systematic geomorphology, as selected by Davis (1903, p. 745–748; 1912, p. 166; 1924, p. 320) and commonly figured in textbooks, is one of the Basin Range type which becomes fringed with a bahada as it is dissected.

Contrasting climatic conditions.—Cursory inspection of many scarps at the outcrops of recently active faults in New Zealand shows that under humid conditions consequent rivers along the bases of fault scarps are by no means rare. In this region numerous rivers in such positions are vigorous enough to carry away the waste derived from degradation of scarps progressively as it is supplied to them. These conditions cannot be called specialized; nor are they peculiar to New Zealand. Even in California, R. Willis (1925, p. 662) recognizes such "longitudinal consequents", some of them in "valleys along single fault lines." Anderson (1941, p. 152) notes that in western Idaho there is "fault-block topography" different "from that of the Basin and Range province in that the basins are not those of interior drainage and are not characterized by the presence of bahadas and pediments."

In New Zealand:

> Most of the large relief features are tectonic forms—of course modified by erosion to a greater or less extent—while the river courses are very largely consequent, still following very closely the courses taken upon the tumbled and irregular surface produced by a late disorderly uplift. In many cases the blocks or units of the disorderly tumbled crust are bounded by faults. (Cotton, 1916a, p. 60)

"Tilted blocks are of very common occurrence, and many of the consequent valleys of larger rivers lie in the fault angles between the inclined surfaces of tilted blocks and the fault scarps of their neighbors" (Cotton, 1922, p. 3).

Examples selected from such rivers which follow courses in places hugging the bases of conspicuous fault scarps are the Awatere (Figs. 1, 6) and the Hutt River, at Wellington

(Pl. 1, fig. 1). These and other consequent rivers undercut the fault scarps in some places, developing river-cut valley-side bluffs; but their most important function is to carry off the debris of scarp erosion as fast as it is supplied to them.

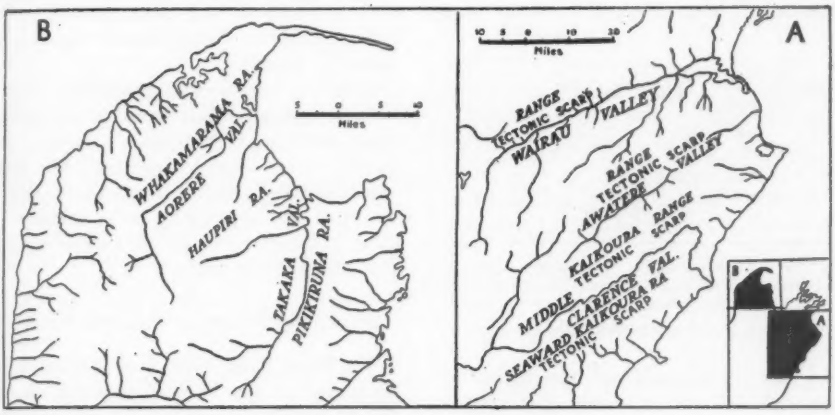

FIGURE 1.—LOCALITY MAPS

A, Eastern Marlborough; B, Northern Nelson. The inset map shows the localities in the South Island of New Zealand.

Dissecting streams on fault scarps.—Every scarp formed by faulting is dissected to some extent by streams consequent on its infantile slope, which will commonly be in part a talus slope of landslide debris, if the movements of faulting, necessarily intermittent, follow one another at short intervals, so that the growth of the scarp is rapid. Talus will be particularly abundant in the case of scarps formed by reverse faulting.

If the catchment area of dissecting streams is confined to the infantile slope, dissection of a scarp of resistant rock must be very slow. Numerous scarps up to 2000 feet high in the South Island of New Zealand retain a steep, little dissected form owing to a delay, or postponement, of dissection which may be thus explained. Waimate scarp (Cotton, 1948b, p. 422) has this wall-like aspect. Dissecting streams may, however, grow headward into an upland or highland in the hinterland of the scarp, and may push back the divide separating them from consequent drainage of weaker gradient on a back slope (assuming the simple tilt-block case). Streams dissecting a scarp can thus become in part obsequent (Davis, 1913a, p. 202). In the case of a block with the crestline initially farther back, perhaps a block with a warped or arched surface, the dissecting streams on the scarp are longer and yet may be wholly consequent. As such streams are well provided with water from the larger catchment area, they proceed at once to dissect the scarp thoroughly, so that it is possible to explain the fully mature dissection of many scarps without assuming that much time has elapsed since the scarp emerged.

Davis (1924, p. 321; 1930b, p. 22) has recognized that some antecedent streams may survive, cutting gorges through fault blocks and scarps. Streams inherited from a prefaulting cycle (Davis, 1913a, p. 194) may exhibit a pattern adjusted to structure, or subsequents may develop during the dissection of a scarp (Davis, 1924, fig. 8, p. 322–323). Some Australian observers have recorded little in the way of development of new consequent streams dissecting scarps they attribute to faulting, though such scarps are said to become zones "of youthful topography" (Süssmilch, 1909, p. 345). Some zones so described in Australia are perhaps not fault scarps, but certain of them may be tectonic scarps developed by monoclinal flexure of the surface. These zones

characterized by relatively deep dissection have been described as exhibiting rejuvenation of streams already flowing prior to the deformation. In the case of actual fault scarps it is obviously possible for some such streams to be extended so as to debouch across the faults,

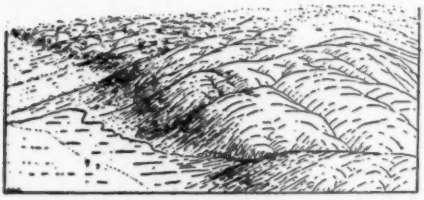

FIGURE 2.—MATURELY DISSECTED ROWSLEY FAULT AND BRISBANE RANGE TILTED BLOCK, VICTORIA, AUSTRALIA
From a photograph

and their presence will lead to rapid and very thorough dissection of a scarp. Notwithstanding such "advanced dissection", however, "ridges all terminate . . . along a piedmont line of simple pattern" (Davis, 1924, p. 321). This is the condition of the northern part of the Rowsley fault scarp, in Victoria, Australia (Fenner, 1918, fig. 30, p. 278), where the Werribee, Parwan, and Lerderberg rivers are involved. The same fault scarp in its southward extension where it bounds the Brisbane Range, a gently back-tilted block, is in a much better state of preservation (Fig. 2), because it is there dissected only by consequent streams and their insequent tributaries.

Spur-end facets.—In the general case of a consequently dissected scarp, the spurs, be they short or long, that descend between dissecting ravines or valleys end bluntly in line at the fault, and in many cases their ends assume the form of triangular facets. Commonly, however, the slopes of these facets do not differ materially in steepness from average slopes in the dissected part of the range behind them.

Complementary relation of fault scarp and bahada.—Fault scarps which have not vigorous rivers evacuating waste along their bases are characteristically fringed by bahadas, but it is recognized that not every fault scarp is so fringed (Blackwelder, 1928, p. 308); the presence of a bahada is a feature of the Basin Range type. At the stage of maturity the fans which coalesce in the bahada are continued headward by aggradation in the valleys dissecting the scarp, as in the case of the mature Inyo fault scarp, in California. Spur-ends between these are not only buried to some extent but are protected by the bahada from extensive river erosion.

Scarp and bahada are complementary. Some bahadas in Australia have been interpreted as relict, and from their presence a former alluviation under pluvial conditions has been inferred (Browne, 1945 p. xi); but these might be explained as growing independently of climatic change merely as the complements of the scarps they border.

Prefaulting Structure of Dislocated Terrains

The Basin Range type of fault scarp is independent of the internal structure of the faulted terrain. Oregon and Wasatch varieties of scarp may, however, be recognized. The former may be defined as characterized by a simple, unfolded structure of strata or lava sheets (as in Oregon) in the scarp-bounded blocks, these being inclined only to the extent that the fault blocks are tilted. If softer strata underlie a cap rock, such scarps when eroded retreat as structural escarpments. Distinct from these are the scarps of the Wasatch variety like those of Utah, in which massive rocks or strata with strongly folded and perhaps faulted structure are exposed. This classification agrees with Penck's (1894, 2, p. 361) subdivision of *Schollengebirge* (block mountains) into *Deckschollengebirge* and *Grundschollengebirge.*

Some textbooks convey an erroneous impression that fault scarps originate only as a result of dislocation of terrains of horizontally disposed lava sheets and stratified formations. Davis (1898, fig. 101, p. 161), who figured an idealized group of the block mountains of southern Oregon as examples, thought at first of fault blocks as consisting "of thick layers of rock, the surface of the uppermost layer forming the back slope". In later years other textbook writers, reproducing slightly altered versions of Davis's Oregon diagram, have indicated the presence of what appear

to be not merely lava sheets but in the main limestone and other stratified formations in the dislocated terrain, these being shown in attitudes which indicate that they have remained horizontal until tilted by block faulting. The popular misconception thus perpetuated of a

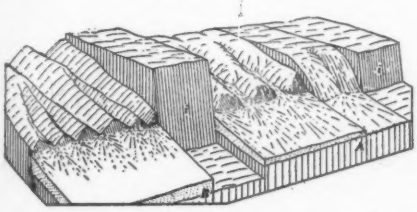

FIGURE 3.—GENERALIZED DIAGRAM OF FAULT SCARPS OF BASIN RANGE TYPE

A, faulting has produced a scarp (of potential initial form, *i*) since degraded without renewal of faulting.

B, growth of the scarp, still continuing, is accompanied by its degradation; the potential initial form (*i*), if erosion were inhibited, is shown, indicating the measure of the displacement.

horizontal prefaulting attitude of the strata in all standard examples of fault scarps is a belated restatement of Powell's (1876, p. 16) original idea of "Basin Range structure," as he named it, which envisaged range-forming faulted blocks "tilted so that the strata dip", the blocks thus becoming "monoclinal ridges of displacement." Davis (1901) soon realized, however, that, as Spurr (1901) had already pointed out, terrains of heterogeneous nature consisting in part of granite masses and in part of old folded welts of strata are predominantly present in the block mountains and fault scarps of parts of the Great Basin.

As Davis (1901, p. 457) has remarked: "In the absence of statement to the contrary it has been sometimes supposed that the faults by which the blocks are limited were determined by ordinary stratigraphic evidence." Assuming that such evidence would in reality be rarely available, he investigated how geomorphic criteria of block faulting might be used, and out of this analysis of geomorphic processes emerged the earliest block diagram in which the presence of diversely folded and anciently faulted strata in upfaulted mountain blocks is taken into account (Davis, 1901, p. 458; *reproduced* Cotton, 1948b, fig. 337, p. 439), as in Penck's conception of *Grundschollengebirge*. Davis shows that there may be a strong presumption of faulting at the boundary of such a block or range, which becomes practically a certainty if, as in many known examples, the folded "structure of the ranges is . . . oblique to their border, so that the faulted margin passes indifferently from one structure to another" (Davis, 1901, p. 458).

Faults bounding blocks in the Basin Range region are commonly normal, but the Basin Range fault scarp is a geomorphic type independent of the known or suspected hade of the fault beneath it. The implication of association with a complementary bahada restricts the type so as to exclude the commonest tectonic scarps of humid regions. Figure 3, A shows this conventionalized scarp and bahada.

Progressively Rejuvenated Fault Scarps

As Davis (1903, p. 745, 749) has pointed out, and Johnson (1929, p. 362) has emphasized, a common variant of the Basin Range type of fault scarp, with spur-end facets very sharp-edged (Pl. 1, fig. 2), and without aggradational fill in the dissecting ravines between facets, has been produced when growth of scarps has been gradual, the growth accompanying progressive, though no doubt intermittent, faulting, which has continued to or almost to the present day (Fig. 3, B).

Many such progressively rejuvenated scarps are fringed by extensive bahadas; but progressive sinking of the foreground instead of upgrowth of the scarp has in some cases been the reason for the absence of aggradation in the dissecting ravines. In such a case much of the alluvium derived from scarp degradation is lowered and buried, so that the bahada fringe is superficially small, as in front of the Panamint and Death Valley fault scarps, which Noble (1926, p. 425–428) and Hopper (1947, p. 398) have described, and the Deep Springs scarp (Pl. 1, fig. 2), cited by Blackwelder (1928, p. 297). Davis (1925, fig. 1, p. 390) graphically indicates the great depth of alluvial fill that may be present under one of these inconspicuous bahadas.

A Fault Scarp of Basin Range Type in New Zealand

The Wairau scarp (Cotton, 1947b, p. 87), in New Zealand, is an example of a Basin Range fault scarp of the Wasatch variety. It looks southward over the western part of the Wairau Valley (Fig. 1) in a rather dry (subhumid) rain-shadow area east of an extensive knot of mountains. The Wairau River, of consequent origin in an elongated tectonic basin parallel to the scarp, flows at some distance from it, having been apparently forced back by growth of a bahada, and the foreground consists of recently deposited alluvium of unknown depth. Though maturely dissected farther east, for upwards of 10 miles at its western end the scarp is straight and almost wall-like in appearance, with a height of nearly 4000 feet (Cotton, 1948b, fig. 270). It is in reality dissected by a number of short subequal consequent streams, the valleys of which cut it into facets. These are strikingly flat and coplanar, but not steep. Unlike numerous other fault scarps in New Zealand, this one seems to be still growing up; at its base there is a rejuvenating scarplet, though this has been formed only after a long episode of dissection. The scarp as a whole is a strikingly conspicuous feature, not because it is particularly steep, but because it makes a rectilinear boundary between an upheaved range and a broad tectonic depression, the Wairau Valley.

Fault Scarps of the Awatere Type

Awatere and Clarence Fault Scarps

One of the most conspicuous fault scarps in New Zealand is that which flanks the broad Awatere Valley tectonic furrow (Fig. 1) for the full length of the Awatere River (Pl. 2, fig. 1). This and similar scarps (initiated by either normal or steep reverse faulting) may be referred to an Awatere type (Fig. 4). The high Clarence scarp (Fig. 1) is broadly similar to the Awatere example, and so also, though on a relatively small scale, are parts of the Wellington fault scarp (Cotton, 1912a, p. 257; 1912b, p. 309; 1922a, fig. 169, p. 163; 1942, fig. 179, p. 173; 1948b, fig. 296, p. 397). The Awatere and Clarence scarps are partly of erosional origin, but they are none the less tectonic forms; the Wellington fault has dislocated a land surface, so that this scarp conforms to the strictest definition of a fault scarp. In each case, however, rounded, blunt-ended

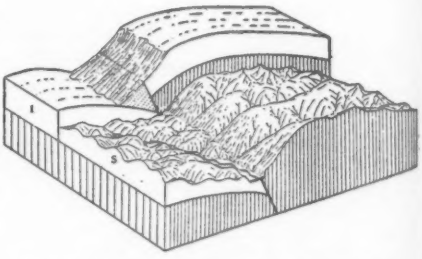

Figure 4.—Fault Scarp of Awatere Type
I, potential infantile form, consisting in part of talus; S, sequential form, as seen on the Awatere fault scarp. Such scarps may be formed by normal as well as by reverse faulting. The relation shown of scarp to river at base is that exemplified by the middle course of the Awatere River.

spurs descend between consequent valleys or ravines to a well defined fault line, where they are sharply cut off.

Though the spur-ends of the Clarence scarp are blunt-ended, and some are broadly faceted (Cotton, 1913, p. 234), the dissecting valleys between them are very deeply cut, each occupied by a small but perennial river which extends far back and branches so as to dissect maturely the Kaikoura Range. The more advanced condition of this dissection as compared with that of the Awatere scarp, probably initiated at about the same time, is perhaps attributable in part to greater height of this scarp; but, in addition, probably the initial form of the high Kaikoura Range was rather broadly arched (Cotton, 1913, p. 234), with the crestline of an initial geanticline some distance back in the hinterland (Fig. 5). High up on spurs that descend to the Clarence scarp there are crestline jogs that perhaps indicate the dissected scarp of a second great fault, though this has not been demonstrated and the jogs are possibly related instead to a detail of the anticlinal form of the range (Fig. 5).

Some of the features of these scarps are characteristic of all scarps of tectonic origin

which, like the Awatere and Clarence examples, have been initially high and very steep or potentially overhanging, so that vast quantities of rock have crumbled, slipped, and been washed away from the crestline parts during upheaval and early in the course of sequential development (Cotton, 1913, p. 234).

Distinction of the Awatere from the Basin Range type depends on absence of a bahada fringe, and this can be explained by the presence of a consequent river parallel to and either close beside or not far distant from the base when the scarp has entered on the long-enduring stage of maturity—i.e. after any landslide accumulations, screes, and steep alluvial cones that may have been formed temporarily during rapid growth and in the stage of infancy of the scarp have been washed away. In the present condition of the scarps mentioned, all the debris derived from rapid early wasting has long ago been removed, and newer debris, though in abundant supply, continues to be carried off progressively by active river transportation, whether the base of a scarp is or is not actually washed by the main consequent river in the adjacent complementary tectonic furrow or basin (Fig. 4).

Thus, even at early maturity of the scarp, either normal valley forms or lowlands of dissection, with moderate to low mature relief, are present instead of a bahada along its base. Only rarely, on the other hand, will the landscape in front of a scarp preserve relics of prefaulting forms of sculptural relief not yet destroyed by postfaulting erosion, for growth of any great scarp has been spread as a long succession of small movements over a vast period, and there has necessarily been much contemporary erosion.

The drainage pattern in front of the base of a high and especially of a potentially overhanging scarp like the Clarence example may result from superposition on bedrock of consequent courses taken across an extensive glacis of debris which fringes the infantile scarp while it is emerging and progressively crumbling (Cotton, 1913, p. 238–240; 1922a, fig. 134b; 1938, p. 195; 1942, fig. 141b, p. 141). It is possible also for a strip of the foreground to undergo pedimentation, such as has developed the extensive Maniototo Plain, in southern New Zealand (Cotton, 1939, p. 231–232). It may be argued, however, that scarp foregrounds of this latter kind pertain only to the Basin Range type of fault scarp, as they testify to the former existence of a scarp-fringing bahada since removed by erosion.

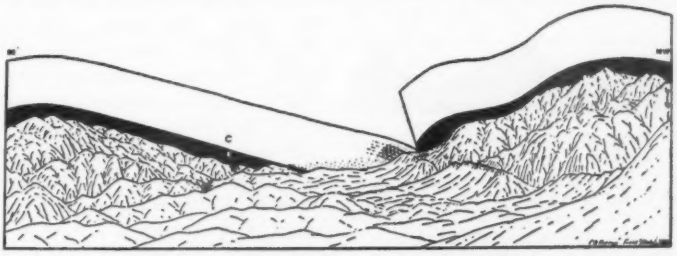

FIGURE 5.—TECTONIC FORM OF CLARENCE VALLEY (CENTER) AND KAIKOURA RANGE (RIGHT)
A generalized section at rear shows the compound structure of the terrain, with (U) undermass and (C) covering strata.

Kaikoura Scarp

Another long and high scarp (Cotton, 1948b, fig. 210) very similar to the Clarence and in a comparable stage of erosional development and dissection to the Clarence and Awatere scarps is that on the parallel Kaikoura fault (Fig. 6), which bounds the Seaward Kaikoura Range on the southeast (seaward) side (Fig. 1, A). These three are a related group of tectonic scarps (Fig. 6), trending northeast, which sharply cut off the anticlinal tectonic forms of three parallel ranges. Unlike the other two, however, the Kaikoura scarp has not a river flowing along any part of its base, and part

of it is fringed by an extensive bahada. This borders the sea, and marine processes are here less efficient than rivers to remove the debris of vigorous mountain-front erosion, so that a fringe of confluent deltas built of bouldery detritus forms a piedmont plain.

FIGURE 6.—GENERALIZED STRUCTURE OF EASTERN MARLBOROUGH

Faults and scarps: A, Awatere; C, Clarence; K, Kaikoura. The fault angles are in reality synclinal, and the ranges between are old-rock cores of fault-broken overturned anticlines. (Compare Fig. 5.) The ranges are seen thus in profile from Wellington.

THRUST SCARPS

Distinction between High-angle and Low-angle Thrusts

Fault scarps at the outcrops of steeply inclined reverse faults have been termed "thrust scarps" (Kerr and Schenck, 1925, p. 480), but such a distinction is unnecessary. Except in infancy, when there will necessarily be a greater development of talus slopes of debris breaking away from the overhanging initial form as it emerges, there is no essential morphological difference between the scarps of such faults and of normal faults. If a special category of thrust scarps is to be recognized, let it include only those tectonic scarps produced by low-angle thrusting, which do not closely resemble fault scarps as generally defined and are commonly far from rectilinear. Where sheets of hard rock have been rather recently transported by overthrusting so that they have come to overlie weaker formations, the low-angle thrust scarps formed at their margins must in the mature stage of development resemble great structural escarpments, retreating as these do and thus becoming sinuous in ground plan. B. Willis (1928, p. 520) has described the scarps of Mount Gilboa and Mount Carmel in Palestine as distinctly sinuous, and has taken this morphological character as indicating relation to the outcrop of a rather low-angle thrust.

In the case of high-angle thrusts the tracing of a long straight line, as exemplified in the Coast Ranges of California (R. Willis, 1925, p. 658) and in the great scarps of the northeastern part of the South Island of New Zealand (Fig. 1, A), seems to be characteristic, more so than in the normal faulting of the Basin Ranges. In the latter, Davis (1927, p. 669) notes: "the trace . . . departs from a straight line in many concave bights . . . between somewhat pronounced cusps"; Gilbert (1928, p. 22) also remarks on the "crooked trace". The block pattern produced by gravity faulting is notoriously irregular, perhaps explaining such cases and the irregularity observed by Noble (1926, p. 425–428) in the great scarps of southeastern California, where the margins of minor fault blocks making "a roughly zigzag pattern" are integrated in high but somewhat sinuous major scarps.

Features of Thrust Scarps

In its infantile condition the scarp at the outcrop of a low-angle thrust would be smothered in talus debris and cluttered with landslides (Johnson, 1929, fig. 13, p. 370). No examples in a youthful stage are known with certainty, but it has been suggested that some scarps in Africa (Wayland, 1921, p. 253) and elsewhere which superficially appear to be on the outcrops of high-angle normal faults or zones of such faulting have been shaped in reality by shallow, or landslide, faulting at the outcropping edges of great thrusts. Fuller and Waters (1929, p. 223–238) have attempted to show deductively that thrust scarps will never become disguised in this way; but it is doubtful whether this has been demonstrated (Cotton, 1941, p. 245). The theory of thrusting has been applied to the upheaval of the Sierra Nevada of California (Lawson, 1936, p. 1711; 1938), the high eastern scarp of which presents an appearance of step faulting "on a host of minor, closely-spaced faults" (Putnam, 1949, p. 1295) dipping at from 70° to 90°. These Lawson regards as superficial, remarking: "It would be difficult to visualize the locus of a great thrust free from secondary normal faults" (Lawson, 1938, p. 438).

The mature condition of a thrust scarp would resemble the eastern face of the northern Rocky Mountains, whether this is or is not

actually, as Bailey Willis (1902) assumes, a first-cycle feature. Great scarps of this kind have been reported also by B. Willis (1912) in a mature stage of development in the European Alps—the northern face of the Wildstrubel, for example.

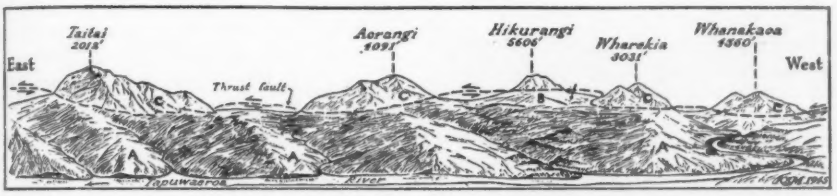

FIGURE 7.—GIANT BUTTES IN THE EAST CAPE DISTRICT, NORTH ISLAND, NEW ZEALAND
Relics of a resistant formation transported on the Taitai thrust. (After E. O. Macpherson, 1946, fig. 1, p. 6.)

In contrast with the more strictly rectilinear fault scarps, thrust scarps fray out and become fringed by large buttes when maturely dissected. Chief Mountain is an outlier thus separated from the overthrust sheet of the northern Rocky Mountains, and, in New Zealand, Mount Hikurangi and other adjacent high buttes in the East Cape district are relics of an overthrust sheet (Fig. 7). Such fraying-out may be entirely the work of erosion on the thrust-faulted structure; but it is possible for strips of a thrust sheet to move forward differentially, separated from one another by flaws, or tear faults.

MONOCLINAL SCARPS

Faults and Monoclines

In the Marlborough district of New Zealand (Figs. 1A, 6) high range-making asymmetric anticlines are cut off by long faults of large displacement, the Awatere, Clarence, and Kaikoura faults. But for the geological proofs of faulting it would remain doubtful whether the scarps on these lines are fault or monoclinal scarps. In many respects they resemble a scarp at Takaka (Pl. 5), in the north of the South Island of New Zealand (Fig. 1, B), which is now interpreted as monoclinal but has been formerly mistaken for a fault scarp (Cotton, 1916a, p. 62).

Takaka Monoclinal Scarp

This scarp faces west, with a remarkably straight base line and a height of 3000 feet, separating the Pikikiruna Range, a recently upheaved mass with an anticlinal core of ancient rocks, from the Takaka Valley, a lowland of synclinal origin; the actual transition slope between these two major topographic units has been shown to be, in part at least and perhaps for the greater part, a steep flexure instead of a fault, though the flexure is broken locally by thrust faulting (Fig. 8, B).

In New Zealand structures developed during the end-Tertiary orogeny, thrust faults replace and form part of the steep flanks of large asymmetric folds; as many so-called fault blocks are really broken anticlines forced up in compression (Fig. 6), this association is quite natural. (*Compare* Washburne, 1928a, p. 175.) Covering strata in the Takaka Valley are not cut off sharply by a continuous fault from the core of Paleozoic rocks forming the Pikikiruna Range, but are upturned, as is shown by a prominent hogback on the outcrop of a steeply inclined bed of Tertiary limestone, which is parallel to the scarp of the range front and close to its foot (Cotton, 1948b, fig. 325). This homocline in the covering strata is not due merely to drag in contact with a major fault, but is a relic of the flank of an anticline which, if restored, passes over the range (Fig. 8, A) (Wellman, 1945, p. 195).

Resemblance of Monoclinal to Fault Scarps

In Figure 9 a monoclinal scarp is compared with a "composite" fault scarp. The former may be sequentially developed either with (*compare* Cotton, 1948b, fig. 325) or without

a parallel homoclinal ridge or ridges projecting from upturned edges of the erosion-bevelled cover. In the absence of such the sequential forms of monoclinal and fault scarps may be indistinguishable from each other, and the more thoroughly the scarps are dissected the

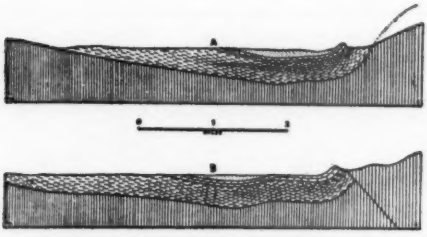

FIGURE 8.—TRUE-SCALE SECTIONS ACROSS TAKAKA SYNCLINAL VALLEY

West of Pikikiruna Range; for locality see Fig. 1. (After H. W. Wellman, 1945, fig. 3, p. 195.)

greater will be the morphological convergence between their forms. A scarp which is submaturely dissected so that blunt spur-ends are ranged in line and which is known by its relation to the structural environment to be a tectonic form is thus not unequivocally a fault scarp, unless the presence of a fault at its foot can be demonstrated. It is permissible to assume, however, that a rigid undermass in a terrain of compound structure will more commonly be broken by faulting than strongly flexed across the ancient structure to make monoclines as sharp as that in the Takaka example. B. Willis (1938a, p. 808), indeed, regarded such flexure as impossible. In the Takaka example, however, the rocks in the core of the anticlinal upheaval include much crystalline limestone, a relatively yielding material when subject to slow deformation.

Australian Monoclinal Scarps

Andrews (1934, pp. 319–325) has clearly pictured the infrequent fault scarps of the east Australian highlands as developing only incidentally and as being subsidiary to warping and broad folding of the land mass and its surface into tectonic forms. The monoclinal eastern scarp of the Blue Mountains is well preserved, with an inclination of from 5° to (exceptionally) 55°, because of the presence of a thick resistant capping stratum that remained horizontal up to the time of the upheaval, but other tectonic scarps in eastern New South Wales are so thoroughly dissected and thus disguised that their restoration can be only doubtfully inferred. Andrews regards

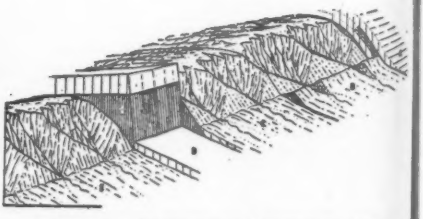

FIGURE 9.—MONOCLINAL COMPARED WITH FAULT-SCARP DEVELOPMENT

A and D, potential initial forms; B, C, sequential forms of monoclinal scarps (with and without homoclinal ridge along base); E, sequential form of composite fault scarp.

the broad valleys of the region as for the most part tectonic furrows, dating from the late-Tertiary upheaval, and their flanks as dissected monoclinal slopes, but where the rocks are less well adapted than those of the Blue Mountains are for preservation of tectonic forms Andrews's thesis lacks demonstration. Some other investigators prefer an alternative explanation of differential erosion, equally difficult to prove (Cotton, 1949b, p. 281).

A line must be drawn somewhere in nomenclature between monoclinal *scarps* and plateau surfaces gently inclined locally as a result of warping. These latter, if extensive and if raised to a sufficient height above sea-level, must in time be dissected, and thus are developed belts of rather strong relief trenched by streams either previously in existence or newly formed as consequents due to tilting. How to distinguish these from scarps descending to pediplains developing in new cycles, general or local (King, 1949, p. 441), is a problem beyond the scope of this discussion. In a plateau region affected by undulatory uplift, as Andrews has called it, there are of necessity strips, broad or narrow, on which the slope has beome steeper. These are strictly monoclinal flexures of the surface, but how steep must they be to be classed as geomorphic

scarps? For the present this also must be left unanswered.

New Zealand Monoclinal Scarps

South and southeast of the Central Otago district of pronounced range-and-basin relief in southern New Zealand (Cotton, 1917a, p. 267–277; 1922, p. 183; 1948a, p. 7–9) there is an area of about 1000 square miles of plateau which has been affected by undulatory uplift, and in this area there are features similar to those postulated as initial forms by Andrews in New South Wales. Here, however, they are in a good state of preservation (Cotton, 1948a, p. 5–7), largely because of the relatively small-scale corrugation, or undulation, the relatively great sharpness of the surface forms it has produced, and the good state of preservation of a well-planed surface truncating schist and graywacke. This is described by Benson (1935, p. 5; 1941, p. 209) as a peneplain of late Tertiary age. Its warped form is mapped by him in generalized contours for the whole eastern part of Otago, but the scale of the mapping is too small to show the numerous minor corrugations (Cotton, 1948a, Pl. 1, p. 2).

The pattern of the undulations of this warped land surface is of finer texture than that of the Australian warpings assumed by Andrews, and is more nearly comparable, indeed, to that found in parts of the California Coast Ranges by R. Willis (1925, p. 648–650), who has noted the presence of numerous strongly arched forms. He remarks that "even in massive rocks the blocks may be deformed by compression, thereby warping their surfaces", which had previously been levelled by peneplanation. Willis has found the profile of the surface arched throughout, or characterized by "lack of inflection points"; it "comes down to the foot of the arch without flattening off at the bottom." (*See also* B. and R. Willis, 1929, p. 336.) He contrasts the convex profile typically so formed with broadly concave forms produced by erosion; it might, however, be thought to resemble Walther Penck's theoretical profile of waxing development.

Monoclinal scarps are common in the undulating-plateau landscape of eastern Otago (mentioned above) just as they are in central and mid-western Otago (Cotton, 1917a, p. 264), where the terrain and surface have been more strongly deformed, and block, or partly anticlinal, ranges are separated by broad complementary lowland basins (Cotton, 1917a, p. 273–277; 1922, p. 183; Macpherson, 1933, p. 270–272). The ranges in Central Otago are arched, but are bounded in parts by well-defined high fault scarps passing lengthwise into monoclinal scarps (Cotton, 1948b, fig. 278, p. 371), or "fold" scarps (Cotton, 1917a, p. 264). On the undulating plateau of the eastern district also, gentler monoclinal slopes of the surface pass in places into definite scarps which, though they closely resemble fault scarps, have not all been shown to be situated on the outcrops of faults (Pl. 3, fig. 1).

FIGURE 10.—WEST-FACING SCARP OF TAPANUI RANGE, OTAGO, NEW ZEALAND
Relief, 2000 feet

The eastern Otago plateau of truncated schist and graywacke slopes down along its eastern margin from an average altitude of 2000 feet nearly to sea-level, dipping below recently deposited alluvium in basins forming part of a tectonic furrow parallel to the southeastern coast. In places the plateau margin is a bold but little dissected scarp, a feature which is conspicuous along the southeastern face of Mount Maungatua, near Dunedin, and is there about 3000 feet high. This part of the plateau margin has been classed both as a fault (Ongley, 1939, p. 23) and a monoclinal scarp (Benson, 1941, fig. 1, p. 210). From the hinterland Maungatua loses the semblance of a mountain and appears as no more than an elongated upswelling of the deformed peneplain forming the interior plateau. Another smooth scarp similar to the steep face of Maungatua faces west towards an asymmetric tectonic furrow, perhaps a fault angle, on the plateau, which guides the consequent south-flowing Pomahaka River. This Tapanui scarp (Fig. 10) is one boundary of an arch or tilted

block forming the Tapanui Range, which stands with a relief of 2000 feet above the depressed part of the plateau in the furrow west of it. Part of the westerly slope appears to be a youthfully dissected monoclinal scarp, but it steepens northward, passing possibly into

FIGURE 11.—UNERODED TECTONIC FORM OF AORERE VALLEY

For locality see Fig. 1. A, Aorere Valley; H, Haupiri Range; W, Wakamarama Range.

a fault scarp which is but little dissected (Pl. 3, fig. 1). Drainage westward from this part of the Tapanui Range is scanty, and the slope therefore retains a wall-like form, merely scored by gullies the catchment of which is confined to the scarp. Unless a suspected fault along its base (Benson, 1935, fig. 2, p. 4) can be demonstrated geologically, the whole scarp, steep as part of it is, must by analogy with the Takaka scarp, as well as with Benson's interpretation of that of Mount Maungatua, be classed as quite possibly the surface expression of a steep monoclinal flexure.

Transition to Fault Scarps

It is obvious that some steep monoclinal scarps pass endwise into fault scarps, as has been suggested in the case of Tapanui scarp. In the North Island of New Zealand, a great fault scarp bounding the Thames Peninsula on its west side is continued southward, through a young splintered fault scarp at Te Aroha, by the smooth westward monoclinal descent of the warped surface of the central volcanic plateau (Bartrum, 1931, p. 5–6). Various other combinations of fault with monoclinal relief are possible; they are recognizable in cases where the combination scarp is still youthful. Scarps combining fault features above and features developed from less steep monoclines below are seen on both sides of the Aorere Valley (Figs. 1, 11), which is as a whole a broad synclinal furrow. That on the northwest side of the valley is marked by the presence of inclined flatiron-shaped terminations which extend the spurs of the valley-wall out beyond the bases of a row of blunt ends or facets aligned on a fault defining the upper part of the scarp (Fig. 12; and Pl. 3, fig. 2). The scarp as a whole descends from the Waka-

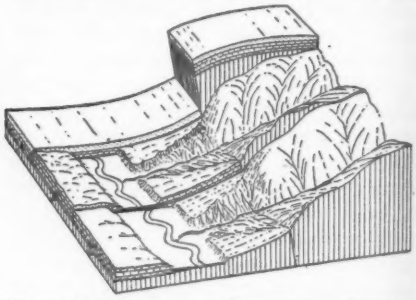

FIGURE 12.—DIAGRAM OF COMBINATION SCARP OF THE AORERE VARIETY

A, potential initial form; B, C, possible sequential forms

marama Range block, or fault-broken anticline, to the floor of a tectonic valley modified only to a small extent by erosion. The base of the combination scarp is washed, because of the synclinal initial form of the Aorere Valley (Fig. 11), not by the main consequent river parallel to it (the Aorere River) but by a relatively weak tributary stream; it has, for this reason perhaps, escaped strong undercutting by river erosion, which might otherwise have destroyed the terminations of the spurs, paring the whole wall back to the fault line or beyond it.

FIRST-CYCLE FAULT-LINE SCARPS STILL TECTONIC FORMS

Fault-line Scarps Developed in the First Post-faulting Cycle

Many fault-line scarps must be included in the category of tectonic landforms, for they mark tectonic, as distinguished from structural, lines and boundaries. These are, in general, scarps rapidly developed by stripping of soft cover from downthrown blocks in the cycle introduced by faulting, as contrasted with those revealed by erosion in second or later cycles along the lines of ancient, long inactive faults. The latter are structural, not tectonic, features (Cotton, 1947d). The explicit definition of fault-line erosion scarps given by Johnson (1929, p.

355) envisages only these structural scarps, but he has also recognized and figured first-cycle development (1929, fig. 3, p. 358). Davis (1913a, p. 206), when he introduced the description "fault-line" seems to have had none but second-cycle forms in mind. Only these could be so named if the ruling of Blackwelder (1928, p. 293) that all fault-line scarps are non-tectonic were accepted; in this case forms which have been classed as fault-line scarps but are first-cycle forms, and which clearly are tectonic, would have to be given some other name or classed simply with well-worn fault scarps, to which they are very closely related.

Forms such as have been classed as first-cycle fault-line scarps are very common in New Zealand (Cotton, 1925, p. 101). Because of the prevailing presence initially of a layer, which becomes in places a very thick cover, of recently emerged and weak strata over a buried undermass offering much greater resistance to erosion, many fault scarps originating on a terrain with this compound structure have very rapidly degenerated either wholly or in part into fault-line scarps (Cotton, 1925, fig. 9; 1942, fig. 188; 1948b, fig. 323; Johnson, 1929, fig. 3), as one must continue to call these forms. In large parts of the New Zealand region, it is this compound terrain that has been broken up by upheaval and deformation into the tectonic arches and blocks that have become major landforms (Cotton, 1945, p. 41). "True" fault scarps are not found as boundaries of these units, if such are to be defined as including only the scarps formed when the worn-down, or perhaps on a lava terrain the initial surface of resistant rocks is dislocated by faulting. (*Compare* Blackwelder, 1928, p. 299–304).

Such restriction of terminology cannot be maintained, however; fault scarps must be made also when a compound (undermass-and-cover) terrain is dislocated. Given sufficient available relief, much, perhaps all, of the weak cover in front of the base lines of these will very soon be removed by erosion, together generally with the whole or most of the cover from block surfaces immediately above the scarps; and the stripping of parts of the faults formerly below ground will develop fault-line scarps. Though it may be said that some such scarps cease to be fault scarps, they are all still tectonic forms.

The Waitaki Valley graben, in southern New Zealand, contains examples of tectonic scarps (Pl. 2, fig. 3) which are also fault-line scarps, though of the first-cycle variety. Within the main graben there is a mosaic of small faulted blocks (Marwick, 1935, p. 332; Cotton, 1948b, fig. 287, p. 381), and the scarps that bound the minor blocks, as well as that forming the northeast wall of the graben followed by the Waitaki River, have been revealed, in part at least, by erosion. Soft Tertiary strata have been stripped away from the downthrown side of faults so as to develop the scarps, in which a resistant undermass of deformed graywacke is now exposed on the upthrown side. An alternative explanation, which must be rejected, is that the block pattern was made by dislocation of a late Tertiary peneplain cut across the graywacke bedrock, such as has been found widely developed so as to truncate all the old rocks of the terrain in adjacent districts (Benson, 1935, p. 5; 1941, p. 209). If this had been the case, the scarps in question would be true fault scarps of the simplest description. Such bedrock truncation could take place, however, only where earlier upheaval had been sufficient to cause removal of all cover, but the presence of thick Tertiary formations in the Waitaki graben shows that the area which later became the graben and its bounding walls was not such a persistent high.

Yet these scarps are tectonic, for erosion has not yet materially changed the appearance of the block pattern of the landscape that was brought into existence by the earth movements responsible for differential upheaval and emplacement of the compound blocks. The pattern, indeed, is strikingly reminiscent of an imaginary fault-line scarp pattern suggested by Blackwelder (1928, p. 299), who has pictured exhumation of a fossil surface "that simulates very closely the forms produced by crustal dislocation" and "exhibits splintered scarps, facets", etc. But the period at which major faulting has taken place is dated later than the latest (Pliocene) peneplanation recognized in adjacent districts. Moreover, the courses of the rivers, Waitaki, Hakataramea,

etc. (Pl. 2, fig. 3), are in the main consequent on the fault-block pattern.

Ephemeral Nature of Fault Scarps in Soft-rock Terrains

Fault scarps must be destroyed by erosion very rapidly when they expose outcrops consisting of soft strata only, and survival of tectonic scarps in such terrains may be expected only in places where large-scale faulting is still exceptionally active. A convincing demonstration of this is afforded the landscape of the Ventura district of Southern California (Putnam, 1942), in which minor forms of tectonic relief are almost entirely unrecognizable, though upheaval has taken place accompanied by strong deformation involving folding and thrust faulting, all of which is extremely recent, for it has followed early Pleistocene marine sedimentation. Such scarps as there mark the lines of faults must, except for rejuvenation on faults that have again become active, be classed as fault-line scarps. Strong tectonic relief as initially developed was almost, if not quite, obliterated in a cycle of erosion, interrupted later, that quickly ran its course into advanced maturity.

If superficial strata are present covering an undermass of superior resistance, the scarp of a fault may disappear during an early stage of the stripping of cover from an upthrown block. Commonly, however, if the cover is not of great thickness, such a scarp is simultaneously replaced by, or may be said to degenerate into, the first-cycle fault-line scarp which eventually emerges if, or when, stripping of cover from the adjacent downthrown block takes place also. The relative magnitudes of fault displacement and thickness of cover determine, of course, whether the degeneration to such a fault-line scarp is complete or whether a basal remnant of the fault scarp which is durable because it exposes the resistant undermass survives above the emerging fault-line scarp. Commonly so little is known of the original thickness of the weak cover, even if a reasonably close estimate of the throw of the fault can be made, that it remains in doubt whether there is any relic of the original fault scarp in the present-day landscape form. Examples thus in doubt are the Waimate and Cave Hill scarps (Cotton, 1948b, figs. 319, 320), in southern New Zealand (Cotton, 1917a, p. 263; 1925, Pl. 1, p. 110), and the Ruakokopatuna scarp in the North Island (Cotton, 1942, fig. 190, p. 181; 1948b, fig. 299, p. 399). Another case of the same kind is the scarp through which the Kern River debouches from the Sierra Nevada of California; this has been described by Blackwelder (1928, p. 297) as entirely a fault-line scarp, but is figured as a fault scarp by Hinds (1943, p. 178–179).

Composite Fault Scarps

Except in drawing one diagram (1924, fig. 9, p. 324), Davis seems never to have deduced the consequences of sequential erosion of those fault scarps which dislocate a landscape with a soft cover overlying hard rock; in this one case he has, by omitting a stage of erosion which exposes a fault-line scarp at the foot of a fault scarp, left the reader with the impression that the development of first-cycle fault-line scarps may be ignored.

There are fault scarps and fault-line scarps developed but recently by degeneration from fault scarps which can be separated only as varieties of the same species. If there is no good reason for separating these more sharply, the distinction may be ignored in some cases also between wholly fault scarps and others which consist of true fault scarps above combined below with scarps exposed by selective erosion. Describing the latter as "composite fault scarps" (Cotton, 1916a, p. 64; 1917a, p. 261; 1925, p. 101) does, however, stress the closer relation to fault than to fault-line forms. Though this case is figured by Johnson (1929, fig. 3, p. 358), he describes it as just one of various ways in which "double" scarps are developed. The chief distinction of geological importance to be made is that between tectonic and non-tectonic forms, the latter produced entirely by selective erosion.

Though some fault-broken parts of Australia have compound structure, Australian geologists have wisely ignored the circumstance that the tectonic scarps there developed, in the Gippsland highlands and Mount Lofty Ranges for example, have become in part fault-line scarps.

Tectonic Scarps Defined

Most of the discussion of fault scarps as tectonic forms is equally applicable to fault-line tectonic scarps, a description which may be used for fault-line scarps developed in a first postfaulting cycle, automatically excluding second-cycle or multi-cycle forms exposed by selective erosion. Blackwelder (1928, p. 293) implies that tectonic, or "true", fault scarps can survive as landforms only where a land surface has been dislocated. Scarps consisting of the unconsolidated sediments covering emerging sea-floors must obviously be ephemeral, but, if the scope of Blackwelder's discussion is widened so as to admit the dislocation of a plain of marine deposition, or "coastal plain", which has beneath the superficial sediments a hard-rock basement, the case exemplified by the Sierra Nevada at the debouchure of the Kern River, tectonic scarps must include some fault-line scarps. It is the practice of most authors to include as tectonic scarps monoclinal and combined monoclinal and fault scarps also.

Recognition of some fault-line scarps as tectonic makes it necessary to emphasize the difference between such (whether or not they form part of composite fault scarps) and the non-tectonic fault-line scarps which appear only in a new cycle after selective erosion has become active along the lines of ancient and dead faults. Davis appears to have had only these in mind when he recommended the adjectival use of "fault-line" in the nomenclature of scarps; other writers have extended its use (Johnson, 1929, fig. 3, p. 358). Second-cycle fault-line scarps must, as Blackwelder insists, be diagnosed and eliminated before identification of tectonic scarps becomes practically possible.

New Zealand Major Relief Largely Tectonic

In New Zealand a profusion of large tectonic scarps, from which examples have been cited, results from the fact that the current major cycle of erosion dates from an end-Tertiary paroxysm of folding and faulting (Cotton, 1916b, p. 248); there was little earlier faulting in Tertiary times on any considerable scale, though relatively weak "orogenies", or spasms of local upheaval and folding, occurred from time to time (Macpherson, 1946, p. 20). An exception to this generalization is the western Otago district, in which an intense orogeny is reported (Benson, 1935, p. 16; Hutton, 1939, p. 86) prior to a regional planation of the surface in later Tertiary times. It seems unlikely, however, that indications will be found of much large-scale faulting contemporary with this in other parts of the region, so as to necessitate a redescription of the great scarps accidenting the landscape as of fault-line erosional and not tectonic origin.

Facet Slopes

Exposed Fault Surfaces

If scarp facets in general were proved to be almost unworn parts of fault surfaces, then it might be assumed that a scarp with well developed facets marked the outcrop of a fault and not the line of a monoclinal flexure. It has been claimed, moreover, by Blackwelder (1928, p. 296) that facets of a little-worn fault surface are nearly as likely to appear on a fault-line scarp as they are to survive on a fault scarp; certainly the fault surface must sometimes emerge rapidly in a process of excavation by selective erosion working under favorable conditions of available relief on soft and incoherent materials in fault contact with resistant rocks.

Both Johnson (1918, fig. 5, p. 235) and Gilbert (1928, Pl. 33, p. 73) have figured a unique slickensided face of basaltic rock (inclined at 49°) in the Klamath Lakes district, Oregon, which is part of a fault surface stripped (in the process of road-making) of a skin of rock debris that has preserved it from destruction by weathering. Johnson describes this as a "rare phenomenon", but Gilbert claims it as evidence that scarps in this region, and by analogy many other fault scarps, present facets which slope with the hade of underlying faults because they are actual parts of fault surfaces.

On the Slate Range scarp (Cotton, 1916a, p. 74), which rises 700 feet above the Gouland Downs plateau, in the mountainous northwestern district of the South Island of New Zealand, there is an isolated facet faced with

a thick slab of quartz, a remnant of a vein in the fissure of the fault that forms the boundary of the range (Fig. 13). Other spurs of the dissected scarp are round-ended or tapering, because of the absence of the resistant quartz, which must have been introduced into the

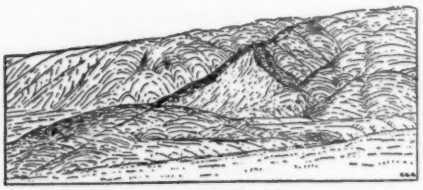

FIGURE 13.—FACET OWING ITS PRESERVATION TO A VENEER OF RESISTANT QUARTZ
Slate Range scarp, Northwest Nelson, New Zealand

fissure prior to the faulting responsible for the tectonic relief, but subsequently to ancient dislocation on the same line. Immediately after the more recent faulting, a slab of quartz has either been exposed in an infantile scarp or, more probably, has been in fault contact with the softer members of a series of covering strata from beneath which the Gouland Downs plateau, now exposed at the foot of the scarp, has been stripped by exhumation (Cotton, 1916a, p. 70). In either case, this isolated and exceptional facet may be of the kind which preserves and exposes the hade of an underlying fault, but this explanation presents some difficulties. Not only must the fault be normal, which seems to be exceptional in this region, where faulting has been associated with compressional folding, but also inclined at the very low angle of about 30°.

Inclination of Fault Surfaces

An investigation by Gilbert (1928, p. 87) has shown the inclination of the fault surface in a fault (or fault-line) scarp breaking the western slope of the Sierra Nevada of California to be nearly 60°, with the faulting normal, though the slope of the facets of the scarp is more gentle. Gilbert, therefore, described these facet slopes as graded, but he was led by field study of the Wasatch scarp, Utah, to reverse his early opinion (1874, p. 50) that Basin Range faults in general are "vertical or nearly vertical," the necessary corollary of which is that their gently sloping facets are graded. He became convinced that the conspicuously flat Wasatch facets, which slope back with an inclination as low in places as 29°, must have been developed, as Davis (1903, p. 754) had already suggested, by survival of a fault surface; he estimated the average dip of this fault plane as 34°.

Gilbert (1913, p. 35) found it necessary to resort to a hypothesis of great crustal stretching in the Basin Range region to account for supposed large-scale movements of normal faulting on such low-angle surfaces. This was followed by the rotating-block theory of the Basin Ranges, introduced and maintained for some years by Davis (1921, p. 92–98; 1925, p. 387–392), though apparently abandoned later (1934, figs. 2, 7, 8, p. 181, 189, 190), which implies that the fault surfaces must be strongly curved with upward concavity to account for much crustal extension without the opening of wide cracks and chasms of kinds not found in natural landscapes. This theory of curved fault surfaces with considerable parts inclined at low angles makes it difficult, if not impossible, to explain why freshly developed basal scarplets, where found close or parallel to submature and mature fault scarps, are generally steep. Davis found it necessary to assume that these mark the outcrops of quite new faults intersecting those responsible for making the original scarps, but it seems more probable that such scarplets are mostly produced by renewed dislocation of the land surface along the preëxisting faults. Even if a scarplet makes its appearance some distance in front of the scarp, perhaps breaking the surface of a bahada so far out that a new line of dislocation is indicated, it is reasonable to suspect that this is merely a minor branch from the main surface of both older and newer dislocation. As Davis (1903, p. 756) himself remarks: "It is to be expected that if a modern fault occurs on a new line it should run across country indifferent to preëxistent structures." Gilbert's suggestion (1928, p. 33–39) that scarplets and minor faulting features in the foreground of the Wasatch scarp are due to shallow gravity faulting in a layer above a ma-

jor low-angle dislocation does not appear to account for such features in general. The theory that fault surfaces sloping back at a low angle have been preserved in little-dissected facets and wall-like faces of young scarps has been applied not only in Utah but has been adopted by both Davis (1925, p. 389) and Noble (1926, p. 425) to explain freshlooking scarps in southeastern California. Here Davis (1925, p. 390) has found it necessary, as in Utah, to postulate movement taking place on steep *new* faults to account for scarplets at the base line.

If older and newer breaks were quite distinct, intersecting faults, it would be a remarkable coincidence if, as is commonly the case, their outcrops were close together and strictly parallel. At Genoa, Nevada, for example, Lawson (1912, Pl. 9, p. 196) has figured a rejuvenating scarplet with an exposed fault plane dipping at 65° which closely hugs the base of high fault-scarp facets sloping back at 24°; and a similar relation of scarplets to facets is found in Californian and New Zealand examples. Assuming that it is true that new breaks in the land surface have been formed generally along steeply inclined surfaces of dislocation, it is then a natural inference that the major faults dip steeply also—in general much more steeply than facets are inclined.

The theory of subsidence along curved surfaces to account for the huge Basin Range blocks seems quite inadmissible. Washburne (1928b, p. 177) claims that all great normal faults are, or have been, active on the upthrow side; his arguments imply that normal-fault surfaces bounding upthrust ranges cannot, therefore, if curved, be concave, but must be convex to the downthrow side. Reviving Gilbert's (1874, p. 50) original postulate of "vertical pressure from below", Bailey Willis (1946, p. 1877, fig. 1), who figures a mechanism involving upward- and downward-moving wedges, argues that major normal faulting will take place on plane surfaces inclined at about 70°.

Origin of Facets

Various observations are on record of dips of faults exposed, some of them in mines, in front of the Wasatch and other ranges in Utah which support the view that the main faults at the range fronts are steeply inclined (Pack, 1926; Gilluly, 1928, p. 1105–1111); these seem to be sufficient to show that the theory of general survival of fault surfaces in facets is a mistaken one. In the Oquirrh Range, Utah, Gilluly (1928, p. 1113) notes that, though well-marked facets slope down to the western fault line bounding the block at only 21° to 30°, the average observed dip of the fault surface is over 60°. It seems highly probable, indeed, that nearly all facets, however suggestive of restoration of a continuous plane surface emerging at the fault line, have undergone considerable change of slope as a result of erosional grading.

Such even grading of the slopes of facets may be simply explained; their slopes do not differ appreciably from other adjacent slopes on rocks of the same kind, and they are thus no exception to the law of equal declivities. These slopes, however, unlike others more irregularly disposed, terminate at the foot on the outcrop of a fault, commonly at a nearly straight line in three-dimensional space. The equal-declivity law and straight base line together explain the back-sloping plane which makes facets. Where this plane terminates a spur, it can produce a sharp-edged facet equally as well as could the exposed surface of a fault.

When alluvium accumulates in front of a Basin Range scarp, burying spur-ends and invading the dissecting valleys, so as to cause the frontal base line to become less regular the scarp passes into the cycle stage of full maturity (Fig. 3, A). Pack (1926), who has noted that bahadas do not bury the bases of the low-angle facets of the Wasatch scarp (Fig. 3, B), regards this as a special case, inferring that along the Wasatch base line "fault displacement has quantitatively kept pace with deposition on the downthrown block." This is the condition that favors preservation of the most perfect facets. Perfect development of facets is not, however, strictly a criterion of youthful dissection.

Facets of Monoclinal Scarps

Graded facets may be developed along a monoclinal scarp (Fig. 9), but steep scarp-making flexures sufficiently regular to define a straight base line are probably rare, though

some are known in New Zealand. On gentler monoclines very broad facets, or flatirons, of a deformed peneplain or exhumed surface may separate bottle-necked dissecting ravines (Fig. 10), but if a monoclinal front is steeper, sloping possibly at 45°, as in the case of the Takaka scarp, the spur-ends may be graded back so as to resemble those of a fault scarp with faceted terminations (Pl. 2, fig. 2). Thus the form of a scarp does not change appreciably where a steep monocline is locally replaced by a thurst fault such as is shown in Figure 8.

Minor Fault Scarps

Deformed Land Surfaces

Certain landforms and segments of the landscape cannot be fitted into any scheme of dissection and selective erosion in the current geomorphic cycle, but must be explained as surviving from an earlier time (Willis and Willis, 1929, p. 335–337). These are not merely the commonly recognized relics of formerly continuous landscapes seen in regions that have been rejuvenated by simple downward shift of base-level; there are other regions characterized by the presence of locally preserved portions of land surfaces, not necessarily peneplains or other nearly plane surfaces, which have been tilted and warped by differential earth movements at the initiation of the current cycle. The landscapes of many such regions are traversed, moreover, by almost innumerable minor tectonic scarps in various stages of progressive obliteration.

Dislocated Landscapes

Identification of minor scarps in dislocated landscapes is very uncertain. So little study has this problem received that it must suffice for the present to confine attention to those scarps which mark the outcrops of faults, such as are present abundantly in seismic regions.

One example of diverse interpretation in such a region is afforded by the Gabilan Range, one of the Coast Ranges of California, all of which are classed by R. Willis (1925, p. 643) as "strictly tectonic", and many of which are bounded by active faults. A group of scarps bounded minor tectonic blocks figured by Kerr and Schenck (1925, fig. 4, p. 479), as well as a general breaking up of the range into minor blocks which these authors recognize, stating that "the same block topography exists in the granite" (p. 481), are ignored in a contemporary account which sums up the range as "a single great fault block" (R. Willis, 1925, p. 667).

This category of minor fault scarps includes not merely the numerous scarplets a few feet only in height formed during known and recent though prehistoric earthquakes, but also many much higher scarps the displacements on which fall into a minor group only because they are dwarfed by great dislocations associated with better known tectonic lines in the same regions.

One law of nature was stated by Isaiah when he announced the principle of peneplanation: "Every mountain and hill shall be made low . . . and the rough places plain"; and the words of the prophecy which immediately precede these may be said to state the law of writhing relief: "Every valley shall be exalted." Throughout the disturbed and deformed regions in which this law prevails there are many faults so recently active that they still break the surface, making more or less conspicuous scarps; but the tectonic nature of these is often overlooked, especially in districts with considerable prefaulting relief, tho sewhich have writhed over a long period with only short intermissions (*Compare* Pl. 4, fig. 1). There fault scarps are easily missed, especially if discontinuous or irregular in ground plan, even though hundreds of feet high. This is true of hard-rock terrains, but where all outcrops are weak, not minor features only but also major fault scarps and tectonic relief in general may escape notice. Erosion, indeed, may almost keep pace with intermittent emergence and growth of scarps, thus relegating them to the category of forms of potential initial relief which have no more than an ephemeral existence.

Exceptionally, however, in regions that are still in the throes of violent deformation, tectonic features may dominate the relief even of soft-rock terrains, as in the case of Boso Peninsula, near Tokyo, Japan (Yamasaki, 1925).

Ideal initial, or infantile, forms of fault scarps associated with considerable prefaulting relief have been sketched by Davis (1913a, fig. 3,

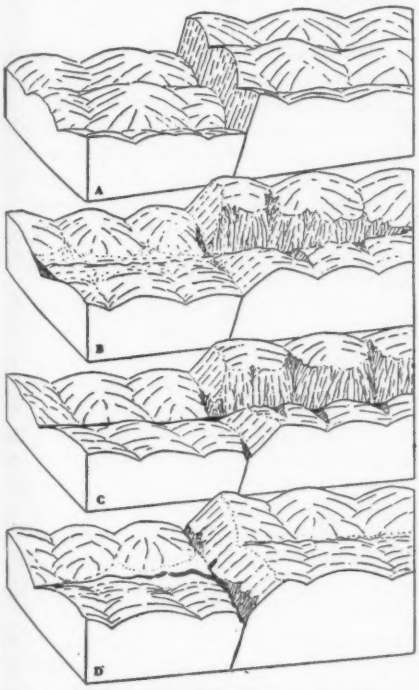

FIGURE 14.—CASES OF MINOR FAULT-SCARP DEVELOPMENT

A, a potential initial form, the effects of landsliding and erosion during intermittent faulting being ignored; B, C, and D show early sequential forms in three selected cases, all different. B, the valley of a river flowing from the upthrown to the downthrown block is incised upstream and aggraded downstream from the fault; the fault scarp is segmented by such valleys, and becomes inconspicuous as a result. C, a river crossing the fault line, flowing from downthrown to upthrown block, maintains its course as an antecedent; if there are several such, again the fault scarp is segmented. D, rapid faulting upheaval causes defeat and diversion of a stream on the downthrown block, and an unbroken fault scarp remains prominent for some time.

p. 194), and some early sequential forms of the landscape after minor faulting are roughly drawn in Figure 14.

Examples of such landscape faulting affecting a landscape of strong relief in North Tyrol are reported by Brückner (1923). Davis (1924, p. 318), agreeing with the tectonic interpretation, has commented that "the faulting appears to have continued into the current cycle of erosion, for certain fault scarps measuring as much as 1000 feet in height have been traced across . . . new valleys."

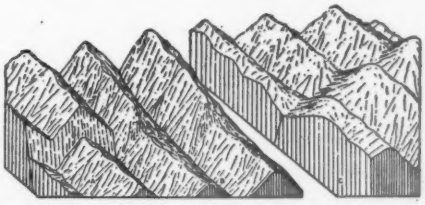

FIGURE 15.—DEVELOPMENT OF CRESTLINE JOGS AND NOTCHES FROM FAULT SCARPS

A, potential initial form; B, crestline jogs developed; C, similar development of notches.

In the case of faults which guide rivers, features usually diagnostic of scarps associated with faults are liable to be destroyed with startling rapidity owing to their mutilation by river erosion. In the train of tilting and warping of the kind to which parts of all strongly seismic regions are subject, great changes of landscape form rapidly succeed one another in river valleys as these are subjected to aggradation, degradation, and terracing; where affected by such conditions the sequential forms assumed by fault scarps will bear little resemblance to standard types as usually recognized.

Crestline Jogs and Notches

The only traces of surface dislocation by faulting that remain in certain landscapes after denudation to a stage more advanced than that shown in Figure 14 are jogs and notches in the crestlines of ridges or spurs. Jogs or notches, according to the direction of downthrow, remain in the landscape after erosion has eliminated all geomorphic traces of faulting in adjacent valleys; on resistant terrains they may survive into full or late maturity of dissection, whatever the prefaulting form of the surface or its postfaulting history (Fig. 15). Notwithstanding enormous erosion since deformation, there are large aligned crestline jogs on the upper parts of great spurs of the Kaikoura Range above the Clarence

fault scarp (Fig. 5); these may be due either to undulating monoclinal uplift or to step faulting on a major scale.

Landscapes with very different initial forms may be broken by aligned jogs in the spur or ridge crests, for example as in Figure 16, C,

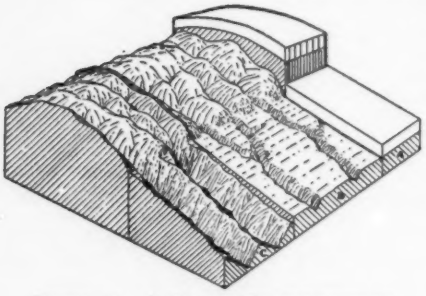

FIGURE 16.—A CASE OF DEVELOPMENT OF CRESTLINE JOGS

Predicted under postulated future conditions. A, potential initial form of slope from Haupiri Range to Aorere Valley (Fig. 11); B, present condition; C, possible future development, with reduction of the fault scarp to crestline jogs on mature spurs.

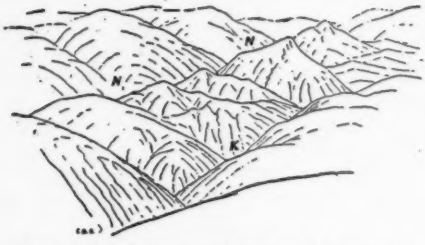

FIGURE 17.—CRESTLINE NOTCHES MARKING A FAULT SCARP

View looking south along the middle Kern canyon, California. K, Kern canyon; N, N, aligned notches in the spurs east of the canyon. (From a photograph by A. C. Lawson, 1906, Pl. 41A, p. 402.)

a hypothetical development following some future upheaval or lowering of base-level sufficient to cause dissection of the present-day landscape (B) (Cotton, 1916a, figs. 5, 6, p. 66, 67), itself derived from a potential initial form A. The vertical fault shown in Figure 16 between the Haupiri Range and the tectonic Aorere Valley is, in the local miners' terminology, the "main slide" (Hector, 1892, p. xv).

Figure 15, C, shows crestline notches developed subsequently to faulting of a mature landscape with downthrow in the general upstream direction. Remarkable notches in spurs descending from the Sierra Nevada upland on the east side of the middle Kern Canyon, California, which isolate large and prominent semi-detached spur-end knobs, have been explained by Lawson (1906, p. 405–408) as due to the presence of a fault which has thus dislocated the landscape. This interpretation is supported by alignment of the detaching notches for some miles, apparently on a rectilinear fault parallel to that which determines the line of the Kern River (Fig. 17. *Compare* Fig. 15, C).

In some cases even isolated jogs may mark the outcrops of short faults, which are commoner in some districts than long rectilinear major faults. Features of this kind are not as rare as the paucity of published references to them might lead one to suppose. Such forms are not, however, in themselves diagnostic of faulting during the current cycle of erosion; similar jogs and notches may obviously be developed in multicycle fashion as fault-line or purely structural erosion features at points where ancient fault lines cross ridges, to say nothing of those structure-controlled jogs, notches, and saw-tooth profiles unrelated to faults which appear in the mature stages of dissection of some stratified terrains.

WORN, FADED, AND MUTILATED FAULT SCARPS

Definition of a Scarp

An ageing and fading tectonic scarp loses something of its appeal to the eye as a topographic or geographic feature but nothing of its interest as a possible indicator of hidden structure. The "essential attribute" of a "scarp" may be, as Blackwelder (1928, p. 293) claims, an "abruptness" which ageing tectonic scarps have lost, but, apart from dictionary definitions, "fault scarp" has long been recognized as a technical term applied to a landform which may be abrupt while as yet unworn, but mellows with age and fades only gradually out of the landscape. Recognition of well-worn tectonic scarps presents greater difficulties than are encountered in the case

of fresh examples, but when possible it is no less important for geological purposes.

Dissected Scarps on Some Major Faults

The Wairau scarp, described earlier as a typical fault scarp, is known to mark the outcrop of a major fault. High tectonic scarps elsewhere in New Zealand which, by comparison with recognized American examples, are less obviously identified by their topographic forms, but which are related in a similar way to geologically known major faults, are recognizable as tectonic because they are the boundaries of prominent ranges. Whatever were the initial and infantile forms of scarps on these lines, notably the Awatere and Clarence scarps, they have been very much worn down and back by erosion, but the basal relics of them are still thousands of feet high and are imposing landforms. It is not because of a close resemblance in topographic development to any standard type that they have been classed as tectonic, but the conclusion that they are the boundaries of great arched and tilted blocks in the tectonic framework of the region, reached partly by a general consideration of geomorphic evidence, is supported by all that is known of the underlying geological structure.

In South Australia, a conservative attitude towards purely geomorphic evidence has been adopted by Sprigg (1945; 1946), no doubt because of the uncertainty inherent in the problem of recognizing fault scarps when they are dissected to full maturity. He has not, therefore, relied entirely on field recognition of such mature features, but has made use of statistical treatment of topographic map data to determine major breaks in the landscape and so define the block pattern. He points out, moreover, that, although the block-faulting interpretation of the Mount Lofty Ranges is "dependent to a large extent upon obvious breaks in topography" (Sprigg, 1945, Pl. 20 and fig. 6, p. 286; *reproduced*, Cotton, 1948b, fig. 282, p. 375), "normal fault phenomena" also are found in close association with the scarps, and "in several instances washouts and quarries . . . have exposed the exact fault line." Scarps must, he almost insists, have associated with them "stratigraphic and/or other evidence of faulting" before they can be admitted as indicating dislocation of the land surface. Geomorphic evidence has, however, proved indispensable even here for distinguishing between the faults which are so recent that they break the surface and other faults more ancient than and truncated by an eroded surface.

Uncertain Interpretation of Some Australian Scarps

In southeastern Australia fault scarps and tectonic landforms in general are recognized by Hills (1934; 1940) only in these districts in which the geological proofs of late-Tertiary to post-Tertiary dislocation are indisputable because of the presence of Tertiary covering strata. In the large parts of Australia which are entirely without such cover fault scarps may have been either overlooked or mistakenly interpreted as fault-line forms, sometimes owing to the conservative attitude of investigators, but more generally because of lack of development of diagnostic geomorphic criteria. In southern New South Wales, when Süssmilch (1909, p. 336–344) diagnosed extensive landscape faulting, he assumed the existence of a number of major faults merely because of an appearance as of dislocation of a peneplain assumed previously continuous over the region; but the evidence for this has since been seriously questioned. (*Compare* Cotton, 1949b, p. 281.)

Interrelating Facts of Seismicity and Tectonic Relief

Interpretation of a landscape relief as tectonic or residual may depend more on what is known already of local geology and geological history than on anything else. Great as may be the theoretical difficulty, however, of formulating rules for distinguishing between fault and fault-line scarps, there is no justification for assuming that tectonic scarps must be rare in an actually seismic region like the middle districts of New Zealand; and conversely, the obviously tectonic major relief of the range-and-basin districts farther south in New Zealand is an argument for a former seismicity much more intense than that of today.

Fault Scarps Mutilated by River Erosion

The presence of faults is not proof that scarps along their outcrops are tectonic, if the faults are no longer active, but in some such cases the presumption is very strong in favor of tectonic origin, even if Blackwelder's test

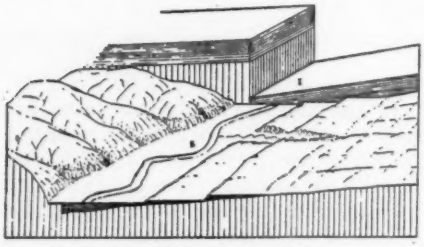

FIGURE 18.—FAULT-ANGLE VALLEY OF A LONG-LASTING KIND

Formed by dislocation of a plain or plane-eroded (perhaps buried) surface of resistant rocks, as shown at rear (I) in a diagram of the potential initial form. S, late sequential form, with valley and fault scarp mature.

"displacement of an older topographic surface" is not strictly applicable. In New Zealand a number of large rivers are consequent in strongly marked major fault angles, so that the overall landscape form indicates that one side of the valley of the river is a fault scarp. This will not resemble a textbook fault scarp closely, for it has either been mutilated by extensive undercutting or has come otherwise to resemble the wall of a wholly erosional valley, except that tributaries tend to come in at right angles instead of dendritically, as they might otherwise do. These are scarps of the Awatere type. An example, in part a fault scarp but partly monoclinal, which resembles the wall of a deeply eroded valley, is a great maturely dissected slope from the crestline of the Kakanui Range to the Shag Valley, in southern New Zealand (Cotton, 1917a, p. 279; 1917b, p. 429). Apart from association with a major fault (Benson, 1935, fig. 2, p. 4; 1941, Pl. 36, p. 216), absence of a complementary scarp or valley wall of similar form and height facing this one across the Shag River (Cotton, 1922a, fig. 143, p. 141) proclaims the scarp tectonic (Figs. 18, 21).

Terraced Fault Scarps

A long rectilinear fault scarp of quite unconventional appearance forms the south wall of a tectonically depressed area containing the Hanmer Basin Plain, in the South Island of New Zealand (Cotton, 1947c, p. 11). Superficially resembling some scarps mutilated by river erosion subsequently to faulting, this is a fault scarp terraced by a river *during* intermittent faulting. The terraces on such a scarp have no counterparts across the river, and so have been termed "unilateral" (Cotton, 1948b, p. 187) (Pl. 3, fig. 3). Instead of such river-cut terraces, benches of bahada or pediment origin might fringe an intermittently growing scarp (Johnson, 1929, fig. 12, p. 368).

These terraces afford evidence of uplift of one bank only of the Hanmer and Waiau rivers, which flow on the Hanmer Plain close to the base line of the scarp (Fig. 22), one westward and the other eastward, joining and passing by way of an antecedent gorge southward through a range that is actively rising. Breaking the lowest river terrace fringing the scarp is a scarplet (Cotton, 1947c, fig. 4, p. 12), which is interesting as one of the first fault scarplets associated with earthquake-making faulting to be described (McKay, 1890, p. 5). Prior to the formation of the celebrated Mino Owari scarplet in Japan in 1891 such forms

PLATE 1.—FAULT-ANGLE VALLEY AND FAULT SCARP

FIGURE 1. HUTT RIVER, NEW ZEALAND, CONSEQUENT IN A FAULT ANGLE

Fault scarp at right. View looking southwest down the tectonic Hutt Valley. (Photo from N. Z. Ministry of Works.)

FIGURE 2. DEEP SPRINGS FAULT SCARP, CALIFORNIA

Photograph by Eliot Blackwelder

PLATE 2.—TECTONIC SCARPS AND RELIEF

FIGURE 1. AWATERE FAULT SCARP

View in the upper Awatere valley. A scarplet at the base is due to recently renewed faulting. (Photograph by V. C. Browne.)

FIGURE 2. TAKAKA MONOCLINAL SCARP, SHOWING FACETED SPUR-ENDS

FIGURE 3. VIEW ACROSS WAITAKI GRABEN INTO TECTONIC HAKATARAMEA VALLEY

Shows block pattern and tectonic scarps of several fault blocks. (Photograph by V. C. Browne.)

FIGURE 1

FIGURE 2

FAULT-ANGLE VALLEY AND FAULT SCARP

FIGURE 1

FIGURE 2

FIGURE 3

TECTONIC SCARPS AND RELIEF

FIGURE 1

FIGURE 2

FIGURE 3

TECTONIC RELIEF AND TERRACED SCARP

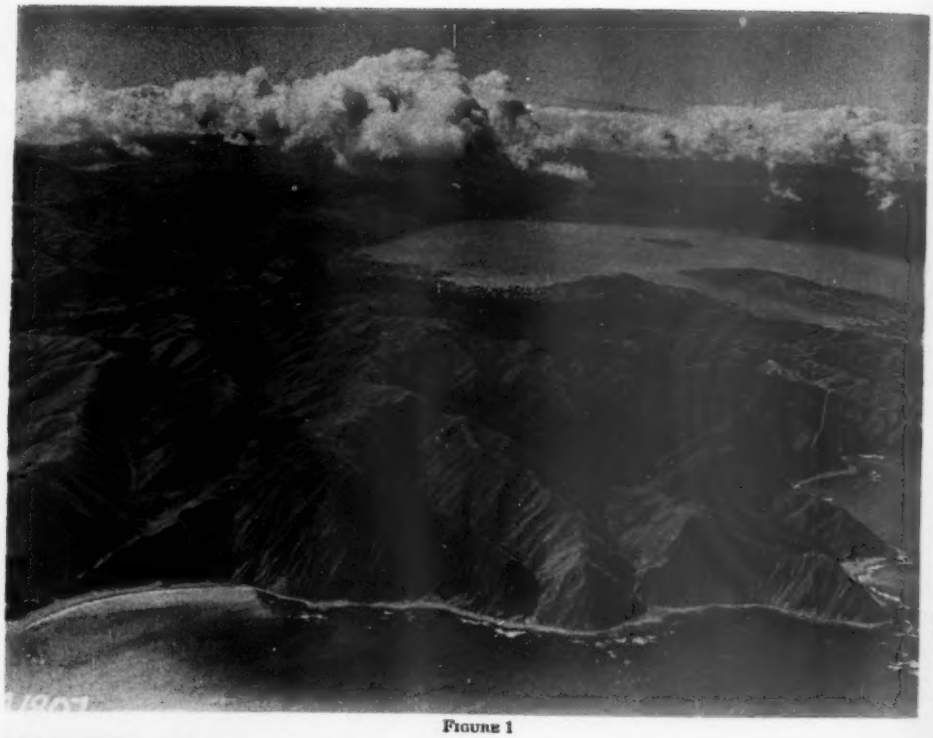

FIGURE 1

FIGURE 2

FAULT-VALLEY DEVELOPMENT

attracted little attention anywhere. Here it affords collateral evidence of the tectonic origin and continued upgrowth of the terraced scarp bordering the Waiau and Hanmer rivers.

There is similar terracing on that part of the fault scarp at Wellington, New Zealand, which borders the Hutt Valley (Pl. 1, fig. 1). Though these terraces are much less conspicuous than those bordering the Hanmer basin plain, and do not mask the conventional fault-scarp form to the same extent, they are distinct gravel-covered benches which are not matched by like forms on the opposite side of the tectonic valley they border. (*Compare* Cotton, 1948b, fig. 340, p. 441.)

Splintered Scarps

Certain scarps that have become the side walls of less strikingly asymmetric valleys than those described above may be recognized as tectonic by the association of minor features which are explicable only in relation to faulting. More or less irregular bench-making splinters such as are well known in association with tectonic scarps may be present and may be clearly distinguishable by their form and their structural relations from terraces developed by erosion at the outcropping edges of resistant strata, while they are too irregular in form and slopes to be ordinary river terraces (Cotton, 1942, p. 244). Obviously, however, such splinter terraces are not positive proof of tectonic as against erosional fault-line scarp, or fault-line valley, origin (Blackwelder, 1928, p. 299). The northern wall of the graben through which the Waitaki River flows, in southern New Zealand, is marked by a number of conspicuous bench-forming minor fault blocks and splinters (Fig. 19) (Cotton, 1917b, p. 432; 1948b, p. 381, 435; Marwick, 1935, p. 332).

A Hawaiian Fault Scarp Recognized by Geomorphic Analysis

A worn and disguised fault scarp of very unusual appearance has been identified by purely geomorphic methods at Waimea, on the island of Oahu, Hawaii (Palmer, 1947). The

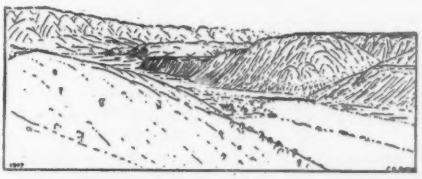

FIGURE 19.—MINOR FAULT BLOCKS BORDERING A GRABEN

The main tectonic scarp on the north side of Waitaki Valley graben, South Island of New Zealand, is broken by cross faulting, with development of small blocks. Flat summit and terrace-like forms are parts of a dislocated surface.

Waimea scarp may be described as a new type, which is peculiar because of its development on the outcrop of a fault aligned parallel to pre-existing river courses which are parallel to one another, though not of fault origin. At Waimea these are radial consequents on a dome-shaped volcanic island. The "scarp" appears in the landscape not as a conventional scarp at all but merely as a pronounced step in the skyline profile of the landscape transverse to the stream courses and fault, though in detail this profile is strongly accidented by ridges and valleys of a maturely dissected terrain (Fig. 20). The origin of the Waimea scarp is indicated by features which, it may be deduced, "would have resulted from faulting and for which no other explanation comes to mind." Among the geomorphic proofs of the fault, according to Palmer (1947, p. 85–89), are: Offsetting both of the adjacent shoreline, where the fault runs

PLATE 3.—TECTONIC RELIEF AND TERRACED SCARP

FIGURE 1. NORTHERN PART OF WEST-FACING SCARP OF TAPANUI RANGE

FIGURE 2. AORERE COMBINATION TECTONIC SCARP
Along the base of the Wakamarama Range

FIGURE 3. TERRACED FAULT SCARP FACING NORTH TOWARDS THE HANMER BASIN PLAIN

PLATE 4.—FAULT-VALLEY DEVELOPMENT

FIGURE 1. VIEW ALONG PART OF WELLINGTON FAULT LINE, NEW ZEALAND
Looking northeastward, at left of center, into the head of Silver Stream fault valley. A fault cicatrice in the axis of the valley indicates activity. (Photo from N. Z. Ministry of Works.)

FIGURE 2. TRENCH-FORM FAULT CICATRICE, RUAHINE RANGE
A scarplet 20 feet high, produced by recent movement, faces inwards towards the fault-scarp front of the range. (Photograph by R. J. Waghorn.)

out to sea, and of the restored smooth contours of the basaltic dome at the fault line, together with the presence of higher wave-cut cliffs at the margin and of two-cycle dissection in the valleys on the upthrown side of the fault. (The upthrow is thus proved to be real, not merely

FIGURE 20.—DIAGRAM OF WAIMEA FAULT SCARP, OAHU ISLAND, HAWAII

Developed parallel to lines of existing drainage arranged radially on a basaltic island dome. (After diagrams by H. Palmer, 1947, figs. 1–6, pp. 86–88.)

relative.) Another point is that, ground water being available in greater abundance at the base of the potential Waimea scarp than on other parts of the porous lava terrain for feeding surface streams, those streams at the fault line flow perennially, whereas others in the vicinity are intermittent. Because of the all-lava terrain, corroborative geological evidence of the presence of the underlying fault is not available in this case. Here, therefore, is a case of strictly geomorphic demonstration of faulting which, it appears, cannot be challenged.

GUIDANCE OF RIVERS BY FAULTS AND FAULTING

The Fault-valley Problem

Some authors follow popular usage and refer to any tectonic hollow or basin as a "valley" (*compare* B. Willis, 1938b, p. 1023), but it is preferable in geomorphology to use the term only for "river" valleys. Davis (1913a, p. 187) has declared that river valleys of "fault" origin are practically nonexistent; R. Willis (1925, p. 659–660), though he has reported fault valleys in California, observes that "they contain no large stream." In New Zealand, however, there are many rivers aligned on faults which are flowing in consequent courses, as is indicated by the relations of such rivers to the boundaries of blocks in a tectonic mosaic. They flow for the most part in fault angles, or fault-angle depressions (Cotton, 1916a, p. 60), a more suitable name for the asymmetric furrows between inclined block surfaces and fault scarps in a block complex than "fault trough," employed by Davis (1913a, p. 194), which suggests a graben.

Consequent Rivers in New Zealand Fault Angles

The Wairoa River, near Auckland, flows for 10 miles in a straight course northward in a

FIGURE 21.—SHAG AND TAIERI FAULT-GUIDED CONSEQUENT RIVERS

Block diagram and map showing the locality, South Island of New Zealand.

fault angle at the base of a west-facing tectonic scarp (Mead, 1930, p. 310; Bartrum, 1939, p. 144). In the South Island, where there are more conspicuous examples, the Shag River flows along the base of the Kakanui Range in a straight course through a major fault-angle depression (Fig. 21). The upper and middle reaches of the Taieri River (Fig. 21), on either side of a horst that forms the Rock and Pillar Range, are similarly placed in the tectonic pattern of Otago.

In the northern part of the South Island (Fig. 1, B), the Aorere River flows in a consequent course parallel to the scarped front of the Wakamarama Range; other rivers, including the Wairau, Awatere, and long middle reach of the Clarence (Fig. 1, A), follow furrows

of large dimensions which are for the most part fault angles, though partly synclinal. In the case of the last two the furrows have been very greatly enlarged by erosion (*compare* Fig. 5), but the consequent character of the rivers in relation to the major features of the landscape is sufficiently obvious to exclude an alternative hypothesis that the Clarence and Awatere valleys owe their development to river erosion working subsequently headward along belts of weak terrain, which, it is true, have been preserved in these valleys because of their downfolded and downfaulted structure (Fig. 6).

Other Rivers on Fault Lines

Besides the many rivers which are clearly flowing in valleys of tectonic origin, some others must be regarded as related to or guided in some unspecified way by faults, though this may be little better than a confession of ignorance as to the mode of origin of their valleys. Commonly it is only the river pattern, as mapped, that suggests control by or relation to a pattern of faults. Valleys are found, for example, which are parallel to one another and not far apart, though it is obvious that they do not owe this relation to subsequent adjustment to the strike of simply arranged inclined or folded strata. Though in some cases the presence of parallel faults can be only inferred from the drainage pattern, in others some faults are either known to be parallel and close to river courses or their presence may be strongly suspected on geological grounds.

The upper reaches of the Taramakau and Ahaura rivers, in the South Island of New Zealand, where they flow due west in straight valleys through the Southern Alps (Fig. 22) are apparently fault-guided (*compare* Henderson, 1937, Pl. 1).

Eastward, across the main divide, the Taramakau fault is continued by one on the Hope River line. Observations, confirmed by photographs, made by Mr. H. W. Wellman (personal communication) show that the gap in this line between the heads of the Taramakau and Hope river valleys (Fig. 22) is filled in by a conspicuously straight reach of the Hurunui River, flowing in an open (glaciated) valley, and that the lineament is further marked by a deep V notch in the ridge between this reach and the head of the fault-guided branch of the Hope. The Kakapo Brook, which, though it is a Hope tributary, runs parallel to the main river for the greater part of its length, as well as other rivers in this mountainous district, form with the Hope an extensive group with

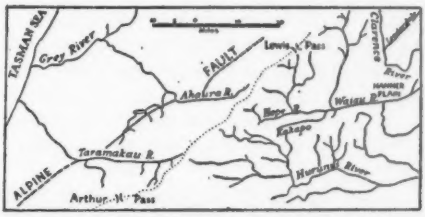

FIGURE 22.—NORTH-CENTRAL PART OF THE SOUTH ISLAND OF NEW ZEALAND

Showing east-west river courses apparently related to a fault pattern.

parallel courses for which a relationship of some kind to a system of parallel faults has been suggested, and seems indeed obvious. Speight (1918, p. 105) has noted the relation of headwater feeders of the Hurunui River to this group. In the mountainous parts of both the southern half of the North Island and the northern half of the South Island there are numerous other examples of similar groupings of faults and fault-guided streams, and there are also in all probability a number of unmapped lone faults which coincide with river courses. Along deep rectilinear landscape furrows through the ranges of this region fault cicatrices betray renewal or continuance of the activity of guiding faults.

In the central part of the South Island, Speight (1928, p. 199–229; 1934, p. 468–474) has noted the presence in the Rakaia and Waimakariri valley systems of a number of valleys and cols that are associated with faults. Though some of the features of the relief are tectonic basins in the sense that they are synclinal forms, grabens, and fault-angle depressions of the land surface, there are also narrower rectilinear valleys that seem to be confined to single lines or narrow zones of faulting. In the case of some of these there is collateral evidence of the presence of faults; in places rivers follow zones of obviously fault-shattered rock; and in a number of valleys of the Southern Alps and outlying ranges there are narrow

downfaulted belts of steeply inclined marine Tertiary strata, nipped in by deformation of uncertain date, which are remnants of formations formerly widespread (Speight, 1914; Mason, 1947). In some valleys, notably that of the Hope River, the nearby Kakapo, and the Leaderdale and other members of the Clarence system farther northeast, fault cicatrices that have been formed obviously in very recent times and may be seen from the air may be followed throughout the length of the valleys and in some cases over low divides from valley to valley, as is the case at the head of the Leaderdale (Fig. 22).

In most of the valleys cited above there is no strongly marked asymmetry of valley form, there is no clear indication that one side of the valley rather than the other is a fault scarp, and the crests of divides on either hand are of about the same altitude. Faults have, however, in some way guided rivers that have taken part in shaping these valleys.

The Fault-line Erosional Explanation

It might be maintained that the only way in which single faults and narrow fault zones can have guided such rivers is by providing zones of crushed and therefore easily eroded rock along which headward erosion accompanied by many captures has by slow degrees concentrated the drainage in fault-guided or "fault-line" valleys, as defined by Davis (1912, p. 170; 1913a, p. 203; 1924, p. 325).

Where faults and their accompanying crush zones are ancient, and where long-continued erosion in one or more cycles has filled in an interval since faulting has ceased, this explanation is the natural one. It has been adopted to explain development of a reticulate pattern of rectilinear valleys in central Sweden by Davis (1912, p. 170; 1913b, p. 519), who has regarded the fault pattern of that region as developed and completed in an age long anterior to that in which valleys now existing have been opened up by erosion, though other authors, notably Nordenskjöld and de Geer (1910), had preferred to explain associated features of the landscape as of modern development in relation to much more recent faulting, i.e. as fault scarps. Such also is the gist of the explanation offered by Marr (1906, p. lxxvii, cii) for some valleys in the Lake District of England—the theory of guidance by "shatter belts."

In New Zealand the fault-line erosional hypothesis has been called on to explain alignment of some valleys near Wellington with faults and crush zones (Cotton, 1914; Gage, 1940, p. 407; Hall, 1946, p. 428), and generally it has been assumed that these fault zones are ancient. Some recently developed fault cicatrices in the vicinity, however, afford evidence of revival or continuance of faulting activity. In one valley, that of the southwest-flowing Ohariu Stream (east branch), which is in some way fault-guided, there has been recent renewal of faulting along a line parallel to the axis of the valley, where a scarplet 3 feet high (Cotton, 1948b, fig. 309) has been formed (Hall, 1946, p. 428). In view of the southeasterly downthrow of this new faulting, though the scarplet is on the southeast side of the valley some distance from the axis, together with the presence of a probable fault scarp forming the northwest valley-side, this valley as a whole may be of fault-angle origin, though it is maturely open and has the appearance of a purely erosional feature. If it is of fault origin, this is probably true also of other near-by valleys with a parallel alignment and generally similar form.

In another part of the same district (Wellington peninsula), a strikingly rectilinear V-shaped but not quite symmetrical furrow continues the line of the Wellington fault scarp (Cotton, 1912a, fig. 2, p. 247; 1922, p. 164), and follows a well-marked zone of dislocation trending southwest diagonally across closely folded strata and ancient strike faults. It cuts notches obliquely through the ridges between valleys of an apparently older system adjusted to a more nearly meridional structural grain. This line of incised narrow valleys, which is occupied in part by the headwater reach of Kaiwarra Stream and, farther southwest, by the upper courses successively of two branches of Silver Stream (Pl. 4, fig. 1), has been attributed to fault-line-guided headward stream erosion along the fault zone (Cotton, 1914, p. 295). In this case, however, the hypothesis of direct fault origin must be preferred, for not only has a strongly marked fault cicatrice been developed in the very axis of the straight V-form

furrow followed by Silver Stream, but there are also on the continuation of the same line clear indications of progressive dislocation, which will be described in a later section.

Hypotheses of Tectonic Origin of Fault-guided Valleys

The rectilinear valleys occupied by the Ahaura and Taramakau rivers (Fig. 22) might be explained either as developed during intermittent strike-slip faulting which perhaps took place during the current major cycle of erosion, or by fault-line erosional development. The first hypothesis is consistent with offsetting at these lines of the Alpine fault (Fig. 22) forming the western boundary of the schist zone of the Southern Alps, which has been mapped by Henderson (1937, Pl. 1). These valleys are incised below the high-level Wainihinihi or Alpine peneplain (Bell and Fraser, 1906, p. 26; Wellman and Willett, 1942, p. 303), a late-Tertiary surface which survives as a crestline bevel on the western slope from the high peaks of the Southern Alps (Cotton, 1947a, Pl. 32, p. 370; 1948b, p. 431) and is not appreciably dislocated at the lines of these valleys. Thus, in the case of the Ahaura and Taramakau fault or fault-line valleys, any hypothesis which implies recent upheaval on one side relatively to the other during formation of a valley is inadmissible.

Along the base of a long northwest-facing scarp marking the line of the Alpine fault (Fig. 22), which is the boundary of the Southern Alps as a structural unit, there is a nearly continuous belt of linear depressions (Cotton, 1948b, fig. 328, p. 431). Though occupied and eroded by many different short streams, the furrow so formed is so nearly continuous for a long distance as to attract attention even on the ground, and has been called the "ancient" Gregory Valley (Morgan, 1908, p. 38). From the air it is very conspicuous. It has been claimed that the valleys on parts of it at least must be subsequent (Morgan, 1908, p. 72; Cotton, 1947a, p. 370). As recent transcurrent movement on the ancient fault line near its southwestern end has been reported (Wellman and Willett, 1942, p. 291), a direct fault explanation of some parts of the furrow is possibly correct, but northeastward there is less clear evidence of this, and both the ancient fault and features aligned on it are mapped as offset by cross faulting at the lines of the Ahaura and Taramakau rivers (Henderson, 1937, Pl. 1).

The references made above to some valleys of possibly fault-line erosional development in New Zealand are sufficient to show that such an explanation has not been established as correct to the exclusion of a direct fault explanation. This emphasizes the need for study of the possibilities of fault-valley development. Alternative hypotheses of fault origin seem to be needed especially to explain valleys aligned on faults, some of them active, in certain New Zealand districts in which the traces of late-Tertiary peneplanation, if ever developed there, have been obliterated by erosion accompanying and following more recent differential movements. Some form of direct fault hypothesis may also explain the incision of valleys along faults even in those districts not so recently redissected after differential writhing, but in such cases it will be difficult to devise tests to establish either a fault or a fault-line erosional explanation.

If a fault hypothesis is to be applied to narrow valleys closely resembling those attributable to unaided river erosion, it may be necessary to postulate earth movements of some special kind; two such fault hypotheses, one of transcurrent movement and another calling for reversal of the sense in which dip-slip displacement takes place, will be proposed later. Tectonic or fault valleys of the fault-angle kind merit description first, however, though they call for no novel theory to explain them; such valleys exhibit considerable variety of form, depending both on the measure and character of the relief of the prefaulting surface and, as has been noted in the discussion of scarps, on postfaulting erosional history. Criteria must be kept in mind for their separation both from fault-line valleys and from such valleys as cannot be explained as consequent in fault angles, though made by faulting.

Fault-Angle and Other Fault Valleys

Tectonic Asymmetry

Large-scale faulting and scarp-making are associated commonly with development of tec-

tonic asymmetry in the landscape; the commonest fault-block forms are tilted blocks, and the basins or valleys between these are fault angles in which the scarp face of one block has opposite to it the inclined back slope of the next. This potential initial form would hold water or guide a consequent stream (Figs. 18, 21).

Davis (1913a, p. 194), though allowing that some "fault valleys" originating in this way are possible, concludes that such must be exceptional, remarking that "when a stream valley is found to follow a fault line it probably has some other than consequent origin" (Davis, 1913a, p. 198). For reasons already given, however, this generalization must be regarded as much too sweeping. In some regions (notably New Zealand) many rivers flow in valleys of this fault-angle kind. A Japanese example is the River Isumi (Yamasaki, 1925, p. 37).

Though the fault-angle concept is a simple generalization, valleys formed as fault angles and fault-broken asymmetric synclines exhibit great variety of form, in part because of the different varieties and stages of sequential erosional development in which they are found, but more especially because of the control exerted by the stage of development and relief the landscape had attained in prefaulting times.

Dislocation of Peneplains

In humid regions, consequent rivers will follow the lines of synclinal downwarpings, grabens, and fault angles produced by deformation of the surface; such forms are conspicuous if a peneplain, or a fossil plain ready to emerge as erosion proceeds from under a weak cover (a case common in New Zealand), is warped and broken. New Zealand examples of rivers following such fault angles are the Shag (Fig. 21), Pareora (Cotton, 1948b, fig. 320) and many others similar to these in southern districts, Ruakokopatuna in the Wairarapa district (Cotton, 1922, fig. 167, p. 161), and Wairoa in the Auckland district (Mead, 1930, map, p. 316).

If the terrain, or an undermass in the terrain near enough to the surface to be resurrected during dissection, is resistant, large fault angles formed by dislocation of a surface of small relief will clearly be stable landforms which will long remain recognizable (Fig. 18).

Besides fault angles, narrow grabens suggest themselves as ideal initial forms of fault valleys, Upper Kern Canyon, California (Lawson, 1904, p. 336), being an example. Few valleys of this kind have been reliably described, but Brouwer (1947, Pl. 4, p. 51) reports them as rather common in Celebes. In broad grabens of symmetrical form, though these become graben valleys, rivers do not commonly follow fault lines closely enough to be classed as fault-guided; asymmetric grabens may be classed with fault-broken synclinal furrows and more regular fault angles.

Dislocation of Hilly Landscapes

If a landscape of mature and moderate to strong relief is faulted, quite strong dislocation may cause little development of new rivers in consequent valleys. Here, however, two distinct possibilities lead to special cases; either new faults may develop or older faults may be rejuvenated. If new faults come into existence, cutting indifferently across the landscape forms already present (Davis, 1913a, p. 190), few consequent valleys will be made at the fault lines (Fig. 14). "Irregular relief along the scarp base will cause any stream that may flow for a short distance along the fault line to depart from it sooner or later and perhaps to leave it altogether" (Davis, 1913a, p. 194). In a thorough analysis of this case, however, many possible variants and complications would have to be considered.

Renewed Faulting on Established Lines

Though river valleys as such cannot guide faulting, some of the valleys already in existence prior to the latest faulting may be assumed to have been fault-guided, being either fault consequents or fault-controlled subsequents. Active faults in many cases are only ancient faults rejuvenated, and thus some recently made fault angles of the surface will be aligned on the valleys of rivers already in existence, as is the Hutt Valley, Wellington, New Zealand (Cotton, 1914, p. 297; 1948b, fig. 340, p. 441) (Pl. 1, fig. 1). In a case like this, the faulting may have downthrow in the original sense or may be

"in reverse" as compared with that which has taken place at the initiation of the earlier cycle. In another example—the formation, or adaptation by recent and still continuing faulting, of part of the Waiau valley, with development of its terraced fault scarp (Pl. 3, fig. 3), as a fault-angle basin containing the Hanmer Plain (Fig. 22)—faulting is quite possibly now in reverse (Cotton, 1947c, p. 17).

Fault-angle valleys or basins thus developed in a landscape previously mature will not be as strongly marked or as long-enduring as easily recognizable features as will the great fault angles produced by large-scale dislocation of a plain or plateau. The large examples cited above are, however, characteristic forms, easily interpreted as long as the fault scarps remain young or submature. In general the valley, which, to simplify the case, may be thought of as hitherto symmetrical in cross profile, is made highly asymmetric by the new fault displacement, one side being steepened by development of a new fault scarp along it.

Many possibilities suggest themselves for further development of such valleys, but it will suffice to select two contrasting cases in which the history is simple. In one of these the post-faulting history—largely indeed contemporaneous with dislocation—has been like that of the Hutt Valley and the basin containing the Hanmer Plain. In each of these fault angles, unilateral terraces now fringe the fault scarp which forms one side of the valley or basin, while lowering of the other side has induced the river flowing in the fault angle to build an aggraded plain, thus embaying valley-side spurs (Pl. 1, fig. 1). (*Compare* Cotton, 1947c, fig. 3, p. 11; 1948b, figs. 225, 298, 340.) In the other case, the streams in the valleys may cut rapidly and deeply so as to incise instead of aggrading them. Such streams are now headwater affluents of the Waipio system, on the Kohala basalt dome of the island of Hawaii (Stearns and Macdonald, 1946, p. 44). These originated as consequents in fault angles originally shallow but converted since by vertical incision into deep canyons, so that the initial asymmetric form of the valleys has been lost.

Fault-angle origin of a main valley may determine not only the overall cross profile of the initial form, but also perhaps the map pattern of tributary junctions. Whereas, on the fault-scarp side, confluences (for the most part of consequent streams) tend to be at right angles, those on the other side may be arranged in some quite different pattern, possibly barbed.

Fault Valleys of other than Fault-angle Origin

It remains to account for the origin of some valleys which are apparently formed in some way by faulting but which are more symmetrical in cross profile than are typical fault-angle valleys and are flanked on either hand by ranges or ridges of roughly the same altitude. A hypothesis of consequent origin as narrow grabens, like the Celebes examples cited by Brouwer, is perhaps not of very general application, since not many deep and narrow grabens are known to exist. Shallow grabens, however, as well as shallow fault angles like the initial valleys cited on the Kohala basalt dome, may have been formed at a high altitude, and courses taken along these may since have been incised deeply.

Another possible valley-making process is successive development of fault scarps which face each other though related to the same fault or fault zone, and this implies that faulting on that line or zone has gone "into reverse." As a working hypothesis this will be discussed later.

Hypothesis of Transcurrent Faulting

Where prominent scarps as well as traces of transcurrent faulting are found on or about the same lines it may not be ascertainable whether the movements that have produced them have been successive or coincident; but in some cases upheaval of a range has taken place in an early faulting episode, while more recent movements have been partly, or perhaps wholly, transcurrent. The actual slip has been a combination of these two components in some minor dislocations of the surface observed to take place during earthquakes in New Zealand, California, Japan, and Formosa (Heck, 1944, p. 212, 216). The enormous dislocation now in progress on the San Andreas fault and its branches in California is, on the other hand, apart from wedging and buckling effects necessarily associated with such movement, wholly transcurrent (B. Willis, 1938a; 1938b; Wallace,

1949, p. 803); but, to judge from meager reports available, primary displacement of the Californian wholly transcurrent kind is uncommon in other regions on a major scale. Some of the San Andreas features seem to be almost unique.

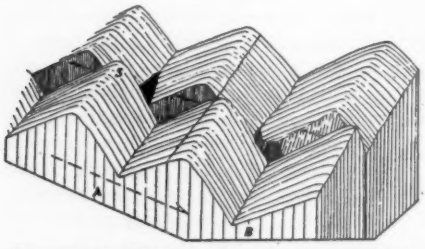

FIGURE 23.—TRANSCURRENT FAULTING, WITH DEVELOPMENT OF SHUTTER RIDGES

S, shutter ridges; A, ephemeral ponding stage; B, stage of offsetting of stream courses.

A theory of tectonic consequent origin, producing a fault valley of a sort, is adopted by R. Willis (1925, p. 659) and B. Willis (1938a, p. 807) to explain parts of the furrow, or "rift," along the San Andreas fault zone, notably in the San Andreas Valley. They regard the valley form as produced by upbulging, apparently in compression, of the land surface on each side of the fault. The rift feature is thus regarded as "the product of two arches one on each side of the fault, formed without vertical displacement on the fault itself" (R. Willis, 1925, p. 659). Production of such a tectonic furrow seems to be peculiar to faulting of the San Andreas type. Parts of the furrow become river courses, but, apart from the San Andreas Valley locality, the San Andreas line is not followed to any extent by streams, "the drainage being across the fault line" (R. Willis, 1925, p. 659) in the main, a generalization verified by vertical photographs (Gutenberg, 1939, fig. 12, p. 205) and maps (Wallace, 1949, Pl. 1, p. 781).

Faults with a transcurrent component of displacement which dislocate surfaces with some relief transect spurs, displacing the spur-ends to form shutter ridges, which, when moved far enough, block the inter-spur valleys or gullies (Buwalda, 1936). Ponding of streams which will take place in valleys thus blocked will be followed immediately by diversions along spill-over courses which will be rapidly incised at the line of the fault (Fig. 23). Offsets thus developed in stream courses will become larger with repetition of transcurrent displacement (Russell, 1926).

The great San Andreas fault is by far the largest well known zone of active transcurrent displacement (B. Willis, 1938a; 1938b), but the topography of this zone throws little light on the valley-making capabilities of such faulting. Apart from the failure of formations on opposite sides of the zone to match, which may be variously interpreted, transcurrent movement in prehistoric but not geologically very ancient times is proved by the offsetting through moderate distances of the courses of many transverse streams (Wallace, 1949, p. 799). Vertical photographs of various parts of the outcrop of this fault show that a belt of offsets aligned on the scar of transcurrent faulting does not necessarily guide a continuous stream for any considerable distance nor become a new valley; but under appropriate conditions it obviously might do so, especially if the movement producing offsets were spread over so long a period that it was accompanied by much contemporary erosion. Where streams are strongly offset, those reaches of offset courses that follow the fault line are fault valleys.

It may be expected, however, that spurs and small valleys will trail over a strip of lowland in the wake of a moving range in which they have their heads. This will be the case especially if intermittent movement has been distributed throughout a zone of faulting, and in such a case the stream and spur pattern will probably degenerate into a tangle which may be irresolvable. A prolonged succession of small transcurrent movements may eventually bring about drainage of such a zone by a single longitudinal river which gradually develops and comes into possession of the zone as its valley. In this process some subsequent (headward) erosion of streams along parts of the distributed-fault zone may cooperate, effecting small captures; but the down-wearing resulting from the integrated work of many small streams crossing a zone crushed and weakened by distributed faulting will be perhaps sufficient to deepen a furrow thus roughed out throughout its length and breadth, provided there is always some way out for water and the waste resulting

from this denudational process. At this stage, trailing of spurs and tributary valleys may develop on either side of the valley, or on both sides, then producing on one side a dendritic and on the other a barbed pattern of stream confluences.

The small Upper Kaiwarra and Silver Stream valleys on the line of the Wellington fault, New Zealand, at its southwestern end (Pl. 4, fig. 1) seem to have originated partly in this way (partly also consequent in a fault angle). The axial streams in these hug the base of a fault scarp along the northwest side, where there is a rectangular pattern of tributary junctions, such as is shown in the background of Figure 24. On the other side, however, there are trailing spurs and tributary streams which suggest the presence there of a zone of distributed movement (Fig. 24). In addition to this evidence of distributed transcurrent faulting in the past, offset features and some shutter ridges prove continuance of lateral movement along a cicatrice which marks a line of active faulting in the valley axis and thence southwestward to the seashore (Pl. 4, fig. 1).

The Opposed-Scarp Fault-Valley Hypothesis

Dip-Slip Faulting Reversible

Nothing is known with certainty of the prime cause of the faulting which makes major landscape blocks of the Basin Range type in the North American Great Basin, but B. Willis (1946, p. 1876–1879) has postulated that the mechanism is "magmatic lift versus gravity". According to this view, "the foundation must be highly plastic," and it is conceivable that processes operating in the subcrustal layer will cause changes in gravimetric gradients such as must lead to displacement of blocks both upward and downward. It seems probable, however, that such a process works continually in the one direction over a very long period; the movements of typical Basin Ranges in Utah have taken place continuously in the same sense, so that recently formed scarplets rejuvenate pre-existing fault scarps, notwithstanding considerable changes in the superficial loading of some low-lying blocks as a result of the temporary presence of Lake Bonneville water over them and the later withdrawal of this load (Gilluly, 1928, p. 1128).

In the Coast Ranges of California and in New Zealand, the deformation of the terrain accompanying fault movements has been of a different kind; in both these regions, as well as elsewhere

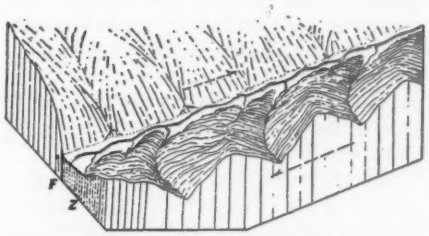

Figure 24.—Stream-Course Pattern Affected by Transcurrent Faulting

On a zone (Z) of distributed transcurrent faulting, a trailing pattern has developed in streams consequent on the slope of a surface tilted towards the base of a fault scarp and tributary to a main stream consequent in a fault angle at the base of a scarp made by dip-slip faulting (F). Streams dissecting the scarp join the main at right angles.

in the Pacific rim, there has been much reverse faulting (Compare Buwalda, 1948, p. 15; fig. 2, p. 16). Lawson (1939) has pictured the ranges of California as an arc similar to those of the East Indies and the western Pacific region, and has developed a hypothesis of eastward underthrusting which might account for upheaval of the Coast Ranges in compression. Others have thought, however, that the compression associated with uprise of these and of the ranges of Southern California may be a secondary effect of the major transcurrent movement on the San Andreas line (Kerr and Schenck, 1925, p. 468; B. Willis, 1938b, p. 1051–1056; 1946, p. 1886).

In New Zealand, the extent to which some end-Tertiary and other still-active compressional faulting (Cotton, 1916a, p. 60; Benson, 1021, p. 64–65; Henderson, 1929, p. 93; 1933, p. 75) is associated with primary transcurrent displacement or alternatively with major underthrusting is not yet known. Macpherson (1946, p. 20) has pointed out that the end-Tertiary mountain-making paroxysm is related to earlier phases of relatively mild folding as the "surface expression of overstressed folds," a range-making episode marking the culmination

of movements spread over a vast period, which intermittently upheaved progressively denuded anticlines in some districts in middle and late Tertiary times. Though the pattern and mechanism were not uniform throughout the region, some of these were embryos of ranges which were later to be thrust up more vigorously with strongly arched form and some faulted boundaries. Much later there has been reversal of the dip-slip movement in some cases; this is indicated by the form of fault cicatrices due to renewed faulting on older lines at the bases of major scarps.

High-angle compressional faulting may easily go into reverse (Clark, 1930, p. 794) and may then be expected to be replaced locally by gravity faulting taking place partly on the same surfaces of displacement. Such movement may supervene when a compressive paroxysm has spent itself. If gravitational adjustment accompanied by some normal faulting occurs, and if crush zones developed by earlier thrusting are still planes of weakness, these will probably guide new faulting along old lines. It has been orally reported that some fault scarplets in the North Island of New Zealand afford evidence of change of the downthrow direction from one observed scarplet to the next visible on the outcrop of the same fault. Such scissors faulting might be merely a secondary buckling effect of transcurrent displacement, but its occurrence is also consistent with a theory that gravity faulting is now in progress.

Geological Evidence of Faults in Reverse

Changing direction of downthrow in successive episodes of faulting is not uncommon. Clark (1930, p. 794) mentions notable reversals in California, and Howard (1937, p. 74) describes an example in Yellowstone Valley. In the case of ancient faults in the Mount Lofty Ranges, South Australia, Sprigg (1946) remarks that, "where the extent of the reversal has exceeded the original throwdown, drag effects suggest faulting in a particular direction, while stratigraphical relations indicate faulting in quite the opposite direction." Woolnough (1920, p. 110) also reports reversal of the sense of faulting in the south of Western Australia, where a block composed of sedimentary strata long ago forced down between upthrust masses of granite has more lately risen again to form the tectonic Stirling Range, structurally a horst (Jutson, 1934, figs. 40, 41, p. 88). A fault scarp recognized by Meynier (1947, p. 170) as forming the long southern wall of the basin of Rennes, in Brittany, affords another example; the faulting which has made this scarp, and is attributed to a pre-Pliocene, or Alpine, orogeny, reverses more ancient faulting on the same line, the surface expression of which has been removed by erosion.

Reversal has taken place on a large scale along some important structural and tectonic lines in the South Island of New Zealand, notably the line of the Shag Valley (Cotton, 1917a, p. 270; 1917b, p. 430; Benson, 1941, p. 216), now marked by a high and prominent scarp (the Kakanui scarp). Along the line of this ancient fault, which has brought into contact relatively lowered graywacke formations to the northeast and relatively raised schists to the southwest, the graywacke terrain has been upheaved again thousands of feet more recently to form the high Kakanui Range, the tectonic scarp of which looks southwestward across the fault over the schist terrain as it is now exposed in the relatively low-lying plateau of eastern Otago (Fig. 21). The more ancient and the relatively recent faulting were separated by an interval, or intervals, long enough to allow of complete destruction of the landscape relief by erosion.

Hypothesis of Opposed Fault Scarps

If the interval between faulting phases has been so short that tectonic relief has not been effaced by erosion before faulting in reverse has set in, the fault scarp formed by the later faulting is faced across the line of fault by the earlier-formed scarp (Fig. 25). Valleys which originate in this way may retain a V-shaped, youthful appearance (Fig. 25), but obviously either such an opposed-scarp fault valley or any other narrow fault or fault-line valley may develop beyond this stage to later maturity, and valleys in any of these categories are subject also to aggradation, degradation commonly accompanied by terracing, and all such complications as are introduced by warping or tilting

in any direction relatively to river flow, especially if they are in a tectonically active district like the northern part of the South Island of New Zealand.

Reverse Scarplets

The hypothesis that opposed-scarp fault valleys are produced by faulting "in reverse" receives support from the occurrence of such faulting now in progress on the lines of some major faults (Cotton, 1947b, p. 17). Certain scarplets forming part of fault cicatrices which have been known in the past in New Zealand as "earthquake rents," but are better called "reverse scarplets," cannot be satisfactorily explained without an assumption of reversal of fault movement. Such a scarplet is commonly accompanied by a trench close beside and parallel to it, or the whole cicatrice, consisting of scarplet and complementary trench, has degenerated into a bench owing to filling of the trench by downhill-creeping debris (Cotton, 1913, fig. 9, p. 235).

There are numerous scarplets, in New Zealand as elsewhere, which merely rejuvenate preexisting scarps, along the base of which faulting is renewed in the original sense; even in the case of the Awatere Valley cicatrice, which forms for many miles a very conspicuous reverse scarplet (Cotton, 1948b, fig. 313, p. 410), scissors faulting causes transition to a rejuvenating scarplet farther southwest along the fault line (Pl. 2, fig. 1). Many reverse scarplets are of such great length, however, as to indicate large-scale reversal of the sense of faulting; an alternative hypothesis of gaping of old fault cracks (Cotton, 1913, p. 237) offers a much less satisfactory explanation of the trench form.

There are few records of such forms outside New Zealand. One, evidently formed very recently, as it dislocates very weak strata, traces the line of a great fault outcropping at the base of the Panamint Valley scarp, in California (Hopper, 1947, p. 399, 426), but this cannot be regarded as a typical example, because strong transcurrent fault movement has been associated with the faulting which originated it. In New Zealand, however, trench-form cicatrices are no rarity, being in some seismic districts characteristic minor relief forms. Photographs of examples in northeastern districts of the South Island have been published by McKay (1902, p. 10, 17), and a number of others in Canterbury, including the "carriage drive" at Lake Coleridge, are figured by Speight (1938, p. 703, 705). Those at the base of the great Awatere and Clarence scarps have long been known (Cotton, 1913, p. 236), and striking examples at the northern end of the Ruahine Range (Pl. 4, fig. 2), in the North Island, are described by Waghorn (1927). In the lower Awatere Valley, where the trench-form cicatrice on the Awatere fault is large and deep, the movement in reverse has been active intermittently in prehistoric and even historic times, and the trench shows signs of development in two at least, but probably many more, earthquake-making jolts (Cotton, 1949a).

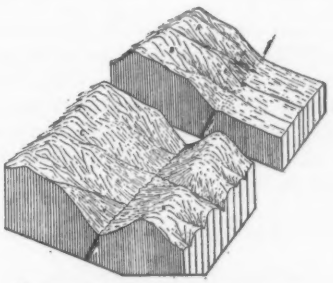

FIGURE 25.—DEVELOPMENT OF AN OPPOSED-SCARP FAULT VALLEY

A, B, C, successive stages. R, newly risen range; V, opposed-scarp fault valley.

True reverse scarplets must be distinguished from features of similar appearance in profile, but short, which are produced quite commonly by movement on both transcurrent and rejuvenating faults where these notch the crestlines of spurs; the geometry of such local scarp development is simple.

The continuous trench to which a true reverse scarplet descends has one of its walls formed by the new scarplet and the other by the steep, straight-sloping facets of the main scarp of a range front. The trench so formed along the main Ruahine cicatrice is rather uniformly 20 feet deep deep for a long distance (Pl. 4, fig. 2). Waghorn (1927, p. 26) remarks:

These westward-facing scarps [scarplets] present an interesting fact in connection with the most recent tectonic movements in the locality. It would appear that the high-standing mountain blocks are now subsiding, the movement taking place, at least in part, along old faults produced when the elevation of these ranges was initiated, or that the area east of the Ruahine Range is now in process of uplift.

Is a reverse scarplet (Fig. 25, B) where found in the landscape of today the premonitor of upgrowth of a new range (R)? If so, and if drainage does not segment this range by cutting antecedent gorges through it, a river may come to flow behind it in an opposed-scarp fault valley, V (Cotton, 1947b, p. 82).

Theory of Oscillatory Earth Movement

As an alternative to the hypothesis of future continuity of movement in one sense after one reversal, it has been suggested that reversal may be temporary and that intermittent movement may be subject to repeated reversal, but only "so that the total displacement evidenced by the uprise of a block mountain is the algebraic sum of many positive and negative movements" (Cotton, 1941, p. 262). This suggestion, inconsistent with the hypothesis of opposed-scarp fault-valley development, was inspired by Gulliver's theory (1899, p. 158) of oscillatory earth movement, or vibration, employed to explain alternating phases of degradation and aggradation due in some cases at least more probably to glacio-eustatic fluctuation of ocean-level. Geologists were not yet familiar with the glacio-eustatic theory and its implications, and so Gulliver assumed a very slow rhythmic oscillation of the land, the ups and downs of which were the prime cause of geomorphic cycles, with smaller and more frequent oscillations marking "epicycles," and minor fluctuations ("vibrations") of which evidence was to be seen in elevated strands and other shoreline features. The whole curve of such changes of level with altitudes plotted against time might, as Gulliver (1899, fig. 2, p. 158) drew it, show oscillation about a mean level, or it might be tilted to indicate a secular movement also in progress.

Not only are such rhythmic vibrations of land level less probable than somewhat similar effects brought about eustatically, but an extension of the vibration hypothesis to relative displacement on faults in explanation of an assumed occasional reversal is inadmissible because of the resistance to any fault movement and especially to reverse-fault movement, which must be offered by the walls of a fault crack. Because of such resistance, fault movement is indeed intermittent, but long-term oscillation or periodic or intermittent reversal, seems to be out of the question (Compare Benioff, 1949, p. 1842–1844; fig. 3).

If the hypothesis of oscillatory displacement on faults be rejected, such reversal as has taken place and is indicated by the presence of reverse scarplets is quite probably permanent; if that is so, intermittent movement which follows must lead to cumulative displacement. Though most known reverse scarplets are low, these appear to have been produced by rather recent renewals (in reverse) of faulting (Cotton, 1947b, p. 81), and it seems probable that some of these scarplets will grow in future to greater height. Some of them have hard and resistant rock on each side; and it is clear that all scarplets are not being worn down as fast as they grow up. The conclusion seems inevitable that in the distant future, perhaps hundreds of thousands of years hence, some of the scarplets will become side walls of river valleys, and presumably similar valley-making has gone on in the past.

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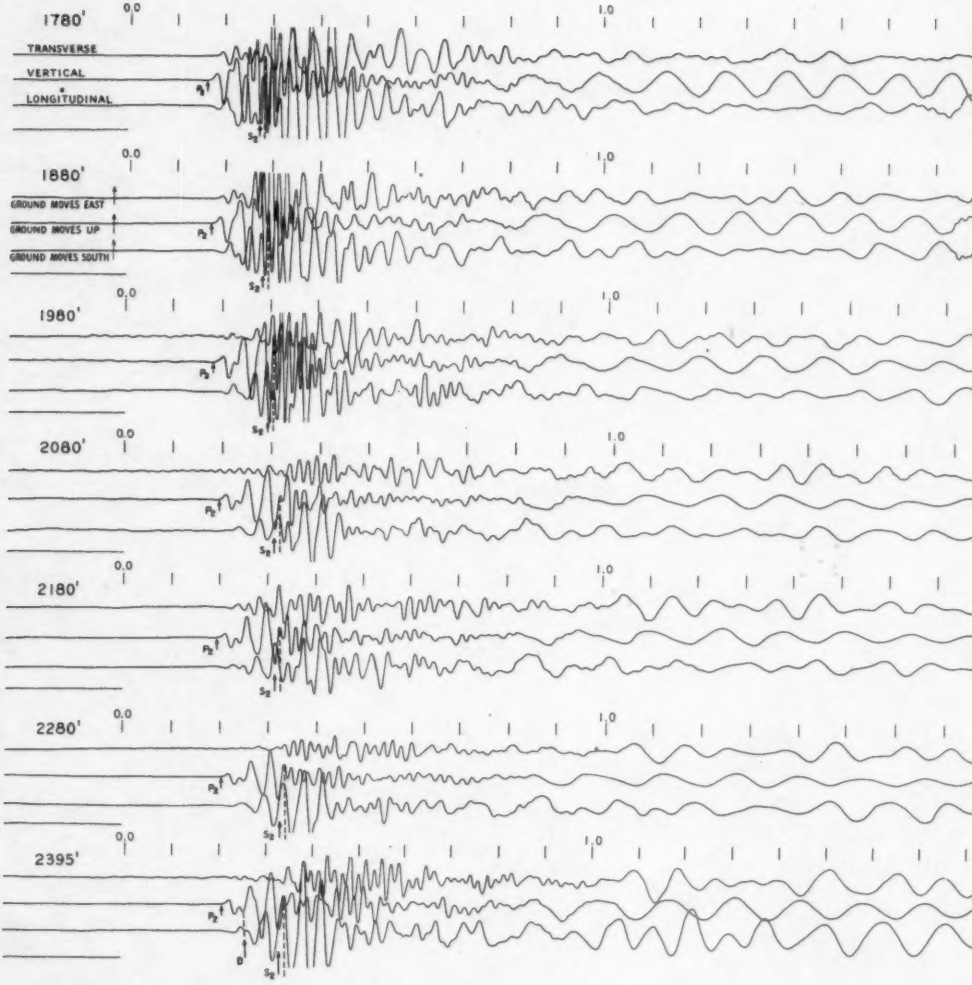
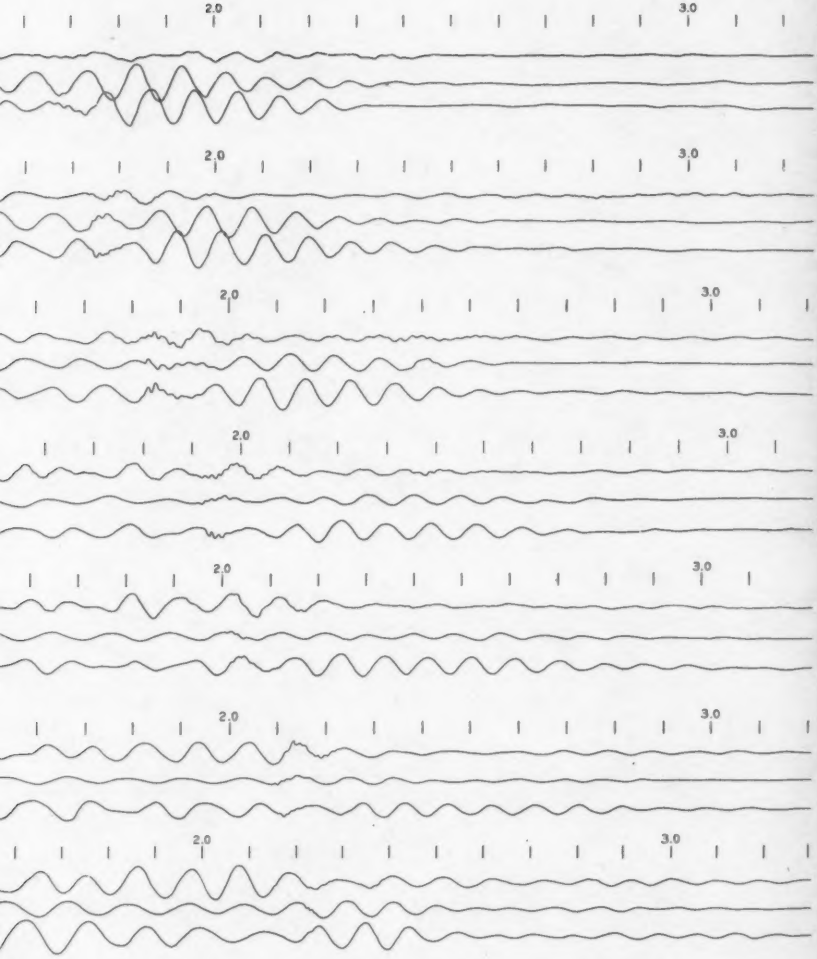


THREE-COMPONENT SEISMOGRAPH RECORDS: WAVES N



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COMPARISON OF FIELD AND LABORATORY MEASUREMENTS OF SEISMIC VELOCITIES IN SEDIMENTARY ROCK

By Charles W. Oliphant

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Abstract

In the field, a small dynamite blast was found satisfactory as the source of both longitudinal and transverse waves. Motion thus generated was detected by a single three-component seismometer, successively positioned at different horizontal or vertical distances from the shot point. With vacuum-tube amplifiers and a galvanometer-type, recording oscillograph, the three principal components of particle motion were recorded on photographic paper as a function of time. Analysis of these records for wave types and wave paths yielded the field observations of velocity.

After the wave path of the field measurements was established, cores from relevant parts of the section were studied in the laboratory. Samples were machined to the form of a right circular cylinder and their resonant frequency for forced vibrations in logitudinal and torsional modes determined. The effect on these resonant frequencies of the variables: water content, pressure, and temperature as well as measurement frequency and amplitude was determined, so that conditions of laboratory velocity measurements could be corrected to those of the emplaced rock.

In Osage County, Oklahoma, four attempts to identify transverse body waves by shooting vertical profiles with a three-component seismometer on the surface and the shot at varying depths in an abandoned hole were unsuccessful. Evidently, the fraction of detonation energy delivered to body shear waves was exceedingly small in comparison with that delivered to other extraneous motions.

Vertical profile shooting was abandoned in favor of horizontal profiles. In Section 2, T.25N., R.4E., Osage County, Oklahoma, velocities of:
$V_P = 9200 \pm 300$ ft/sec. $V_S = 5000 \pm 200$ ft/sec. were measured in an upper layer of Permian sand and shale, approximately 250 feet thick, and:
$V_P = 14{,}000 \pm 400$ ft/sec. $V_S = 9900 \pm 300$ ft/sec. in a lower, inhomogeneous layer of limestone.

After correcting to the most probable conditions of water content, pressure, and temperature for the rock, in place, 17 core specimens from 8 closely spaced horizons in the lower limestone gave a bimodal distribution of laboratory velocities with the modes:

$$12{,}600 < V_P < 14{,}500 \text{ ft/sec.}$$
$$8100 < V_S < 8600 \text{ ft/sec.}$$
$$18{,}400 < V_P < 20{,}400 \text{ ft/sec.}$$
$$9500 < V_S < 10{,}100 \text{ ft/sec.}$$

Conclusions are:

(1) Transverse elastic body waves of the *SV* type can be generated by a small dynamite blast in sedimentary rocks if the shot hole is shallow and high confining pressures are avoided.

(2) The velocity of an *SV* type body wave can be determined with approximately the same accuracy as a longitudinal type body wave. The transmission path is subject to greater uncertainties.

(3) Among the independent variables affecting velocities in a near-surface, Permian limestone, water content is the most important. Pressure, temperature, and measurement direction, amplitude, or frequency have relatively minor effects.

(4) The Neva limestone may be subdivided on the basis of density and laboratory determinations of velocity into two distinct types of alternating microlayers with thickness of the order of 1 foot.

(5) Under the conditions of this work, elastic parameters and Poisson's ratio computed from field measurements differ significantly from values determined by laboratory measurement.

Introduction

Scope of Investigation

To determine the probable composition of the crustal layers and of deeper portions of the earth the velocity of longitudinal and transverse seismic waves has been determined at particular depths from a study of earthquake and quarry-blast seismograms. When these velocities are compared with the values from laboratory measurements on various rocks, minerals, and metals available at the surface, it is reasoned that the elastic parameters of similar materials, under the same environmental conditions, are equal and thus provide a basis for identifying the unseen matter (Jeffreys, 1929; Leet and Ewing, 1932; Leet, 1933; Birch 1937; 1941).

The principal objective of this work is the application of such a method to near-surface, sedimentary rocks which have received less attention in this connection than igneous rocks. When the three basic elements of the method—field measurement of two types of wave velocities, laboratory measurement of these velocities under proper experimental conditions, and the nature of the material—are independently known the validity of the method is tested, and accessory knowledge about the stages in its application are revealed, for example: the conditions of generation by a small dynamite

blast of transverse body waves, the identification of wave types and paths, the applicability of simple elastic theory to sedimentary rocks, and the variables controlling seismic velocities in rock.

Previous Work

In the category of field studies, much work has been done with a mechanical three-component seismograph by Leet. (*See* Leet and Ewing, 1932, and Leet 1933; 1938; 1939; 1946.) Special efforts to study the velocity of both longitudinal and transverse waves generated by a dynamite blast have been made by Leet in the above references, by Ewing, Crary, and Lohse (1934; 1935), by J. A. Sharpe (1942a; 1942b; 1945), by Ricker (1944), by Ricker and Lynn (1950), by Weatherby, Born, and Harding (1934), by Horton (1943), and by Wood and Richter (1931). A falling weight was used as the generating source for the work of Nasu, Hagiwara, and Omote (1936).

In the category of laboratory studies, special mention should be given the work of Ide (1936a; 1936b; 1937), of Birch (1937; 1941; 1943), of Birch and Bancroft (1938a; 1938b; 1940; 1942), and of Iida (1938).

Acknowledgments

The aid and counsel of Professors L. Don Leet and Francis Birch of Harvard University, and of my father, A. G. Oliphant, are sincerely appreciated.

The elements of the three-component seismograph were built at the Engineering Laboratories, Inc., of Tulsa, Oklahoma, under the direction of L. Don Leet and the writer. A. V. Dayton and L. M. Swift contributed valuable design suggestions.

The laboratory study was made in the Dunbar Laboratory of Harvard University with equipment built and used for similar studies on igneous rocks by Birch.

Seven thin sections were prepared at the expense of the Division of Geology and Geography of Harvard University. Professor E. S. Larsen, Jr., and Eugene Robertson very kindly studied them.

Edward Schmitz drafted the illustrations.

Field Study

Characteristics of Seismic Field Equipment

The three-component seismograph consists of three separate channels arranged to produce a record of the three mutually perpendicular components of ground motion at a point; its objective is to describe the motion of this point as a function of time so that wave types and arrival times may be determined.

The seismograph used for this work consisted of a three-component seismometer, three direct-current, vacuum-tube amplifiers, and a recording oscillograph with coil-type galvanometers. The three inertia elements of the seismometer are mounted on a single base with their axes mutually perpendicular; when packaged for field use, the unit weighs 108 pounds and occupies a space approximately 15 by 19 by 10 inches.

The inertia elements of the seismometer were adjusted to similarity, and the following constants apply:

Horizontal element 1:
Damped Period: 0.351 seconds
Per cent Critical Damping: 28.0
Vertical element 2:
Damped Period: 0.350 seconds
Per cent Critical Damping: 28.8
Horizontal element 3:
Damped Period: 0.352 seconds
Per cent Critical Damping: 28.5

The elastic suspension of inertia members was carefully planned to minimize response perpendicular to its axis.

Each of the three direct-current amplifiers was similarly constructed and had identical performance characteristics. Relative gains of each channel were set and locked at values such that trace displacements on the records were of substantially equal value for the same velocity of ground motion exciting each of the three inertia elements. The absolute gain of the three channels was then controlled by an input and output attenuator.

A further refinement included a calibration system which permitted a field check of both relative and absolute gains of each amplifier channel. Selective filters were not employed.

The recording oscillograph employed four critically damped galvanometers with a resonant frequency of 20 cycles per second; one galvanometer recorded the output of each amplifier channel, and a fourth, the shot instant. The Elinvar timing line vibrator was adjusted to a frequency of 50.006 cycles per second by comparison with a standard 1000 cycle signal from Cruft Laboratory of Harvard University, and this same standard was later used in determining the laboratory velocities of small rock samples. Optical parallax and amplifier-phase shift were shown to be negligible by simultaneous application of transient voltages to the inputs of each amplifier channel and directly to a galvanometer of similar type. All seismograms were recorded at a speed of about 15 inches per second on photographic paper 3½ inches in width.

Over-all system performance is indicated by Figure 1 which relates the ratio of recorded trace displacement to ground displacement over a range of frequencies. This curve applies in the special case when ground displacements can be described by the relation:

$$u = a \sin \omega t,$$

and was obtained by combining the calculated response of the inertia member and its voltage output with the measured response of the amplifier and galvanometer.

A further indication of over-all performance is gained from Plates 1, 2, and 3. It should be noted that the particle velocities produced by natural ground unrest equalled or were less than the recording threshold. Thus, waves reflected from deep horizons, and probably many of the transformed waves, should not be prominent on the records.

For all field operations, only one three-component seismometer was available. A separate shot was thus required for each record at varying distances along a profile.

Geology of Area

General statement.—The horizontal profile on which this work is based extends 3016 feet in a N. 9° W. direction in the east half of section 2, T. 25 N., R. 4 E., Osage County, Oklahoma. Six shot holes, approximately 50 feet in depth, were drilled at 600-foot intervals along the line of profile, and a single hole 290 feet deep was located 1223 feet from the north end of the profile. Figure 2 shows the regional and detail location.

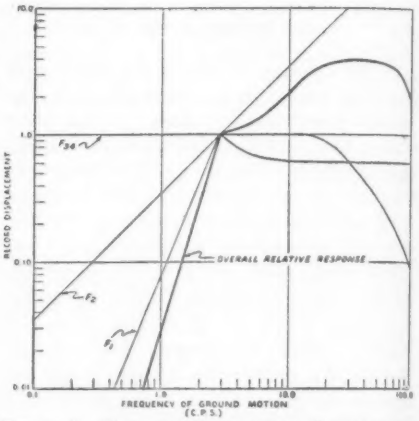

FIGURE 1.—RELATIVE RESPONSE OF SEISMOGRAPH FOR CONSTANT AMPLITUDE OF GROUND MOTION

Ratio of recorded trace displacement to ground displacement as a function of frequency is shown by the heavy line. Lighter lines F_1, F_2, and F_{24} show, respectively, the relative response of the seismometer inertia mass, relative output voltage of the moving-coil seismometer, and relative output voltage of the amplifier combined with galvanometer displacement.

Topography.—Osage County, reservation for the Osage Indian tribe, is administered from the principal town of Pawhuska, 32 miles east of the profile. The western part of the county is underlain by Permian limestones and shales which do not favor tree growth; the area is thus a grass-covered plain with numerous long, narrow limestone benches, many of which parallel the regional strike. Like much of the Osage Plains the topography is mature, and the relief, seldom exceeding 200 feet, reflects the differential erosion of gently dipping limestones and shales. Surface elevations along the profile range from 1066.7 to 1082.7 feet above sea level.

Rock units.—The sedimentary section and corresponding seismic section are shown in Figure 3. Core specimens used for laboratory velocity measurements are described in a later section.

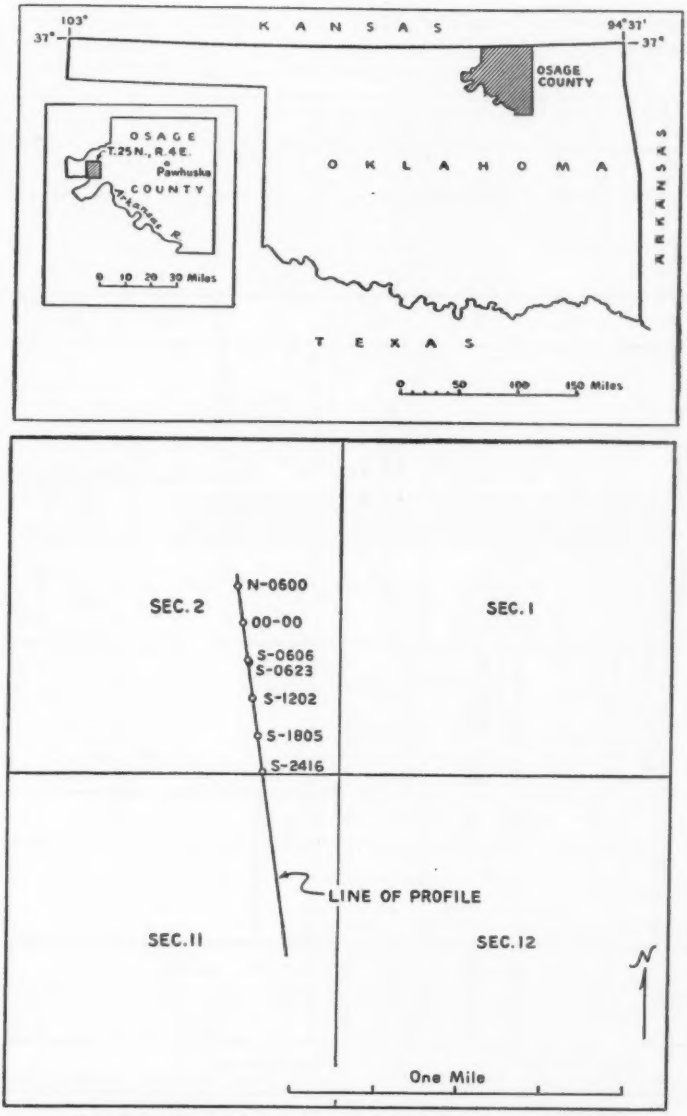

FIGURE 2.—INDEX MAP OF AREA AND SHOT HOLE LOCATION MAP

The columnar section of Figure 3 is compiled from well cuttings and core data to a depth of 290 feet, or slightly below the base of the Neva limestone. Below this depth the section is an interpretation of the electric log from the Helen Bear No. 2, in SE-SE-SE of section 2, T. 25 N., R. 4 E., 800 feet east of the profile. The electric log could not be extended to the shallower beds because they were masked by surface casing.

The electric log is a useful index to the homogeneity of the section since the measured

FIGURE 3.—COMPARISON OF GEOLOGIC AND SEISMIC SECTIONS

self-potential and resistivity should be constant in a homogeneous material.

The velocities and depths in the seismic section are the results obtained from field work.

Classification and naming of the rock units are modified after Heald (1916). In later studies (Moore and Moss, 1934; Moore, 1936; Moore *et al.*, 1944) the entire section has been placed in the Big Blue series.

Structure.—The regional structure, judged by the attitude of the Permian surface rocks, is a homocline with southwesterly dip of 50 feet per mile. Numerous local folds, of interest in the search for petroleum, are developed on the homocline.

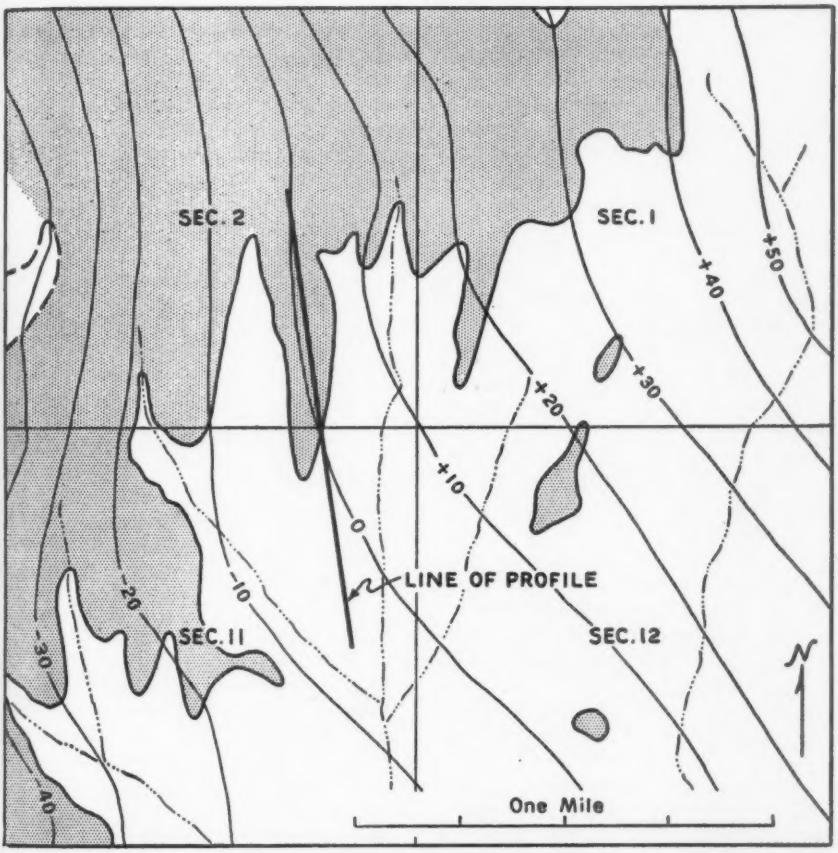

FIGURE 4.—GEOLOGIC STRUCTURE MAP

Structure contours show the elevation on top of the Crouse limestone in T.25N., R.4E., Osage County, Oklahoma. Contour interval: 10 feet; Datum: 1063 feet above sea level. Dotted area is underlain by Crouse limestone. Modified after White (1922).

In the vicinity of the profile, the axis of a gentle fold (Fig. 4) plunges southwest at an angle equal to the homoclinal dip; the horizontal projection of the axis trends approximately N. 45° E., and the axial plane dips steeply southeast.

Along the line of profile the maximum structural relief of the Crouse limestone, known from the core data of six shot holes, is 7.4 feet. The relief of the deeper (248 feet below the

FIGURE 3.—COMPARISON OF GEOLOGIC AND SEISMIC SECTIONS

surface) Neva limestone is undoubtedly no greater, and thus the underlying rocks may be considered horizontal for the seismic velocity measurements; reversed seismic profiles confirm this.

Faults are unknown in the immediate area.

Influence of geologic features on seismic results.—Velocity uncertainties due to dipping beds were eliminated by shooting in a line parallel to the local strike, and corrections for topography and the weathered zone were simplified and reduced to nearly constant small values by shooting in a line with minimum topographic relief.

A further element not easily evaluated in the planning stages concerned the permissible wave paths and the amount of energy transmitted by each of the thin limestone beds. For example, when seismic waves of varying length are incident upon a layer of infinite lateral extent, finite thickness, and elastic parameters differing from that of the material above and below, the amount of energy partitioned among the transmitted, reflected, and refracted waves of the original type and the reflected, and refracted transformed waves will be governed by 12 quantities: the wave length of the incident energy, the angle of incidence, the thickness of the layer, the elastic parameters and density of the material above, in, and below the layer; curvature of the wave front must be considered. Exact relations, however, are unknown because of the complexity of the problem. Particular simplifications by Knott (1899), Sezawa and Kanai (1936), Slichter and Gabriel (1933), Muskat (1933), and Muskat and Meres (1940a; 1940b) are helpful but not definitive for the stratification encountered here.

By experiment, it was found that the thin, 6–9-foot Crouse limestone, and the 12-foot Cottonwood limestone, can transmit energy for short distances, but in very small quantity relative to that of the thicker beds below; the maximum possible sensitivity of the recording equipment was required to observe longitudinal waves, directly transmitted or refracted by these thin, high-velocity layers near the surface. Velocity comparisons were thus attempted only in the deeper Neva limestone.

By reducing the threshold sensitivity of the amplifiers to a level at which normal ground unrest caused less than 1 mm of trace displacement on the seismogram, direct or refracted longitudinal and transverse waves traveling in the thin upper limestones were not recorded. At the lower sensitivity, refraction arrivals from the deeper Neva limestone gave no indication that ray paths had been affected by the presence of thin high-velocity layers above, and on this basis the heterogeneous section above the Neva limestone and below the seismically weathered zone is considered a single low velocity seismic layer, V_1.

Field Procedure

In the earliest field work, problems of generation and identification of transverse body waves were studied on vertical profiles in which the seismometer was placed on the surface of the ground and charges detonated at varying depths beneath it. The brachistochronic path is thus established as a vertical line between charge and seismometer, and directly arriving plane longitudinal waves should produce only vertical components of ground motion, while plane transverse waves should produce horizontal components. Objective identifications of the transverse body wave should be clearly possible.

Seventy-eight seismograms were recorded at four holes ranging in depth from 178 to 3719 feet; each hole contained, for its full length, a column of drilling mud or water which tamped the charges of from ½ to 30 pounds of solidified nitroglycerine or 60 per cent high-velocity dynamite.

Analysis revealed that interval and average velocity values for longitudinal waves were consistent for similar parts of the sedimentary section and agreed well with vertical velocity surveys in surrounding areas. The horizontal traces on the seismogram were very complex, however, and transverse wave arrivals were not clearly distinguishable. A study of the (1) abrupt change in period of horizontal turning points, (2) abrupt change in amplitude of horizontal trace deflections, (3) coincident or in-phase turning points on both horizontal traces, (4) abrupt change in the relative phasing of the two horizontal traces, (5) change in the direction of particle rotation in the hori-

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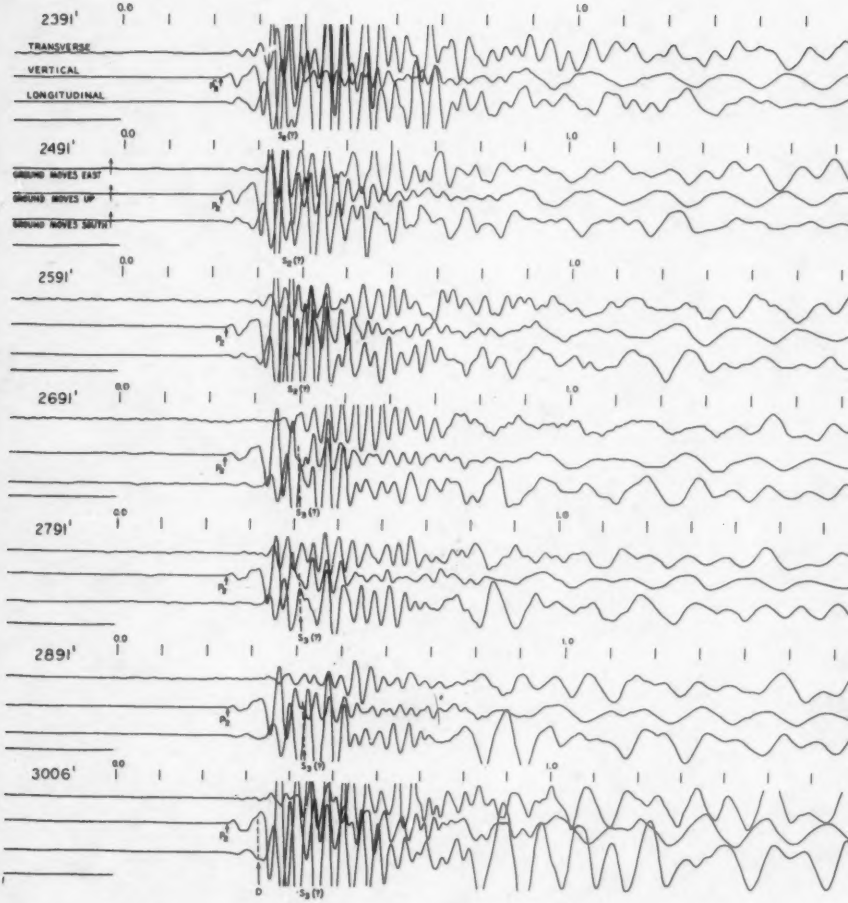
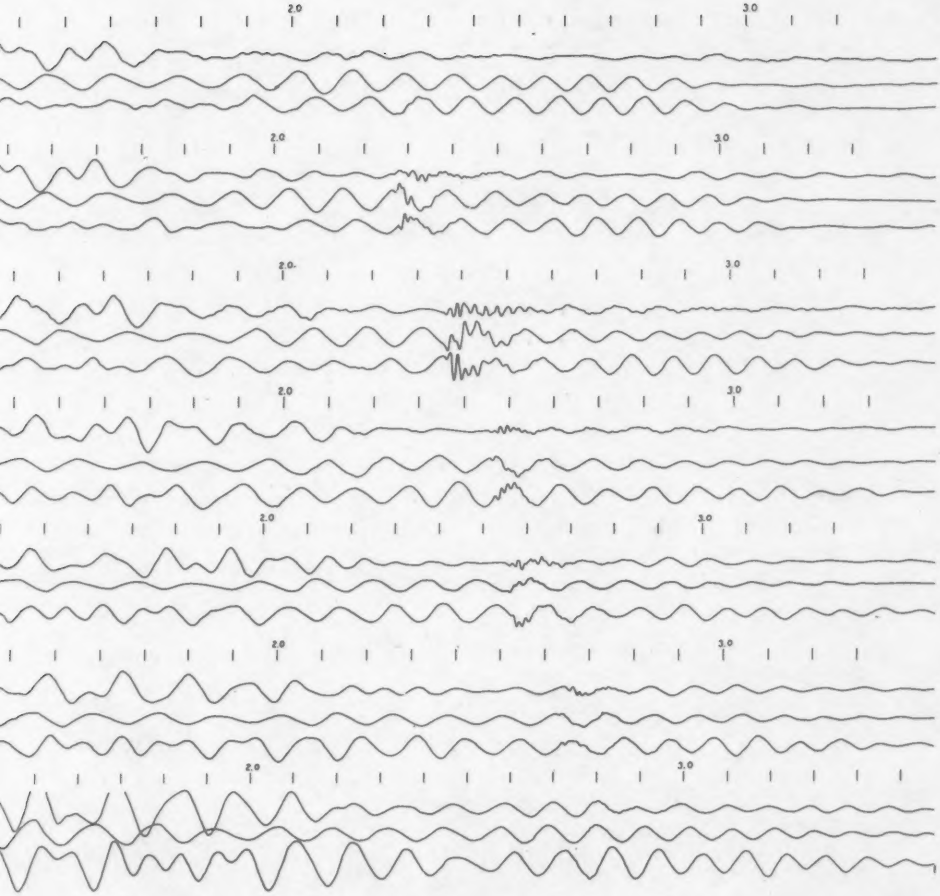


THREE-COMPONENT SEISMOGRAPH RECORDS: W



RDS: WAVES NORTHBOUND FROM SHOT HOLE S-2416

zontal plane, aided in part by a mechanical trace resolution device which computed the direction and magnitude of the resultant of two vectors, showed that these events could not be consistently correlated over appreciable changes in vertical path distance, and thus could not represent the arrival of transverse body waves generated at the shot. At a particular spread distance, recorded events could be duplicated only when no physical element of the system was varied.

The fraction of detonation energy delivered to body shear waves is probably exceedingly small under these conditions, and complex disturbances of the hole itself, the pipe, when present in the hole, and the tamping fluid column probably produce horizontal ground motions of obscuring magnitudes.

Thus, because transverse body waves could not be identified on vertical profile records, the alternative of horizontal profiles was adopted. Successive charges, untamped by fluids or solids, were detonated in shallow holes as the horizontal separation of shot point and seismometer was increased by short intervals along the line of profile. More of the detonation energy is thought to be delivered to transverse body waves under these conditions because of increased opportunities for asymmetrical pressure applications in near-surface, untamped holes.

To measure velocities in the Neva limestone refracting horizon 248 feet below the surface wave paths must be identified, and it was thus necessary to determine, velocities in the two layers above the limestone: the V_0 layer, or seismically weathered zone extending from the surface to depths of 12 to 16 feet, and the V_1 layer extending from the base of the weathered zone to the top of the Neva limestone. A horizontal datum was established within the V_1 layer, and V_2 refraction-arrival times were corrected to this level for changes in surface elevation and weathered zone thickness. For the proper recording sensitivity, the heterogeneous sedimentary section is thus reduced to a simple two-layer refraction problem.

The thickness and longitudinal wave velocity of the V_0 layer were determined at the six shot holes by vertical profiles with a charge detonated at 5-foot intervals of depth. For seismometer locations between the shot holes, the thickness of the V_0 layer was obtained by straight-line interpolation between the nearest holes. Transverse wave velocities in the weathered zone are known with less certainty because of the difficulty in identification at these very short distances. An approximate value was obtained, however, from the transverse wave, V_1, intercept time of short horizontal profiles, when it is assumed that the weathered zone has the same thickness for both longitudinal and transverse waves.

Velocity of the two wave types in the V_1 layer was determined from horizontal refraction profiles over a spread distance of 116 to 716 feet. The seismometer was retained at its initial position while charges were successively detonated in 4-foot auger holes at intervals of 50 feet along the line of profile.

For the V_2 layer, longer spread distances were required, and the untamped charges were detonated below the base of the weathered zone at depths of 16 to 36 feet. In one series of tests the seismometer was bolted to a cement pier based on the Crouse limestone at S-2432; seismograms were obtained at 600-foot intervals by shooting in each shot hole along the profile. A second series of seismograms was obtained in the reverse direction at 100-foot intervals over the horizontal spread distance from 1780 to 3006 feet by placing the seismometer on the ground at 100-foot separation points between shot holes 00-00 and N-0600 and detonating, for each placement of the seismometer, a charge at shot hole S-2416 and S-1805.

One hole was drilled to a depth of 290 feet at S-0623 from which 2-inch-diameter cores of the Neva limestone were secured for laboratory study. This hole was then used for further seismic studies; a horizontal profile was recorded at spread distances up to 1784 feet with all charges at or near the top of Neva, and a vertical profile with charges at 30-foot intervals from the base of Neva to the base of the weathered zone.

Analytical Procedures

Identification of wave paths.—The ray paths traversed by longitudinal and transverse waves

may, for the purposes of this work, be identified by an application of the laws of geometrical optics. The simplification is introduced that all layers are of constant velocity and that Snell's Law is rigorously followed.

To identify the wave path of a significant event on the records by time-distance data (Slichter, 1932) it is important to recognize that events which do not represent the true *first arrival* of energy over some path will result in erroneous path identifications. This is a frequent difficulty when events other than the first recorded motion are studied.

In practice, wave paths are identified by plotting raw, uncorrected data to obtain a first approximation to the velocities and wave path. With this knowledge and the application of appropriate corrections, final identification of the wave path may be accomplished by comparing the travel times, corrected to a suitable datum, with those that would be predicted theoretically from a particular geometrical distribution of the different layers. This comparison is most satisfactorily achieved by noting the deviation of the corrected travel times from the line which predicts their theoretical value.

Identification of wave types.—For this work, interest centers on identifying the true beginning of phases P_1, S_1, P_2, and S_2 which are respectively the direct longitudinal and transverse waves in the V_1 layer, and the refracted longitudinal and transverse waves in the Neva limestone V_2 layer.

Since, by choosing the proper horizontal distance, phases P_1 and P_2 can be recorded as the initial motion on the seismogram, no particular barrier obstructs identification of the true beginning of these phases, hence their wave paths and velocities.

The identification of the true beginning of phases S_1 and S_2 is difficult because they appear as later arrivals on the seismogram and are subject to confusion or obliteration by preceding or simultaneous arrivals. Such complexity often reduces the science of seismology to a subjective art in which the skill and experience of the interpreter plays an inordinately large part.

The most important single criterion for identifying longitudinal or transverse waves is the distinctive motion inherent in each wave type. This motion is best understood by examining the particle displacement of each wave type relative to the same ray path.

In the upper portion of Figure 5 the particle motion relative to a horizontal ray path is shown for plane-longitudinal waves and plane-transverse waves of the *SV* and *SH* type. In the first instance the particle is displaced in a direction *parallel* to the ray path and for the second, along some preferred "line", in the plane *perpendicular* to the ray path. Strictly, the wave may be termed *SV* or *SH* only when the ray path is horizontal and the preferred direction of motion is respectively vertical or horizontal.

In the lower portion of Figure 5 an example of the record character to be expected from plane-longitudinal and transverse waves is shown; the direction of wave travel, wave path, and seismometer orientation were chosen to fit the conditions of actual seismograms in Plates 1–3. Hypothetical pulses of pure longitudinal motion and pure transverse motion of the *SV* type are applied to the above conditions, and separate sketches show the expected record character. In a more realistic case, the motion of the two pulses would be recorded on one record and separated by a time interval governed by path distance and the ratio of longitudinal and transverse velocities.

The sketches emphasize (1) that the largest motion for a steeply emergent longitudinal wave pulse is on the vertical trace, whereas the largest motion for a steeply emergent *SV* pulse is on the horizontal lengthwise (longitudinal) trace, and (2) that the directions of trace displacement are opposite for the longitudinal wave pulse, where a vertical trace peak occurs simultaneously with a lengthwise (longitudinal) trace trough, whereas the directions of trace displacement are the same for the *SV* pulse, where a vertical trace trough occurs simultaneously with a lengthwise (longitudinal) trace trough.

A change in the conditions of the example produces changes in the expected record character; for instance, if all conditions remained the same except that the waves were southbound, the longitudinal pulse would produce a vertical trace peak simultaneously with a

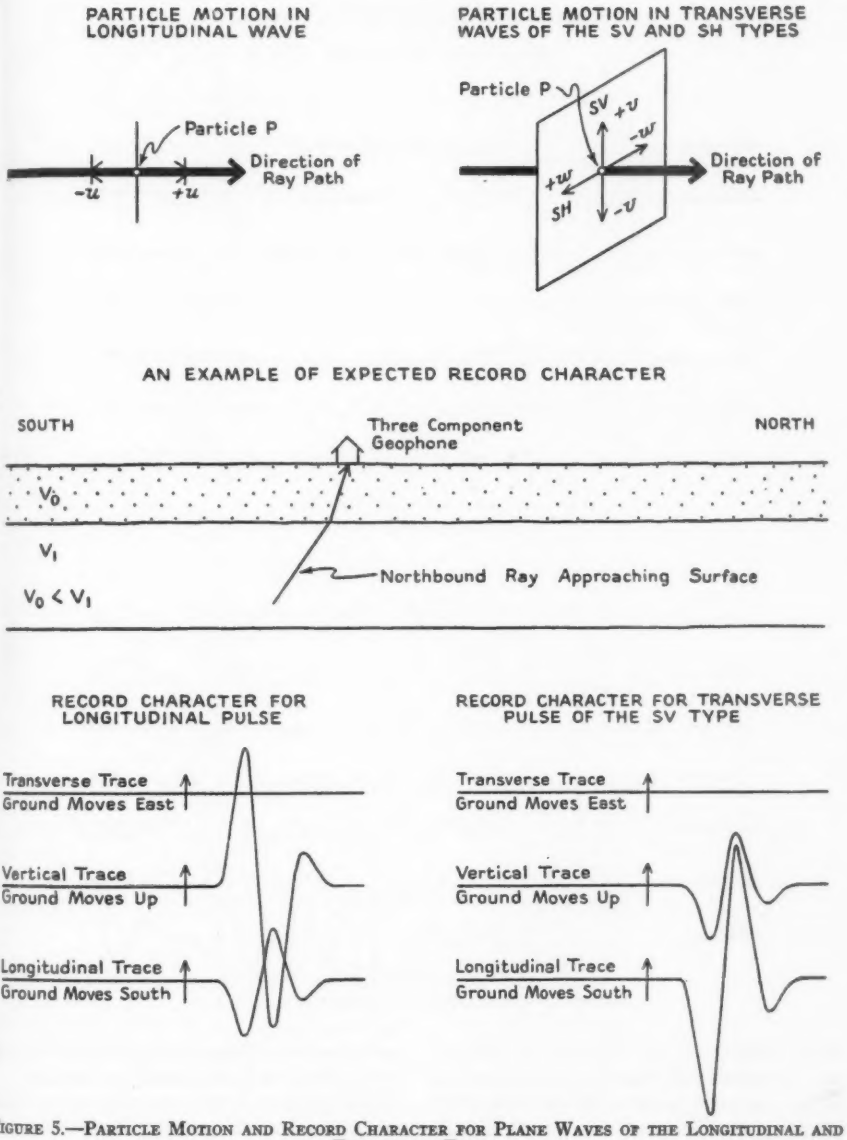


FIGURE 5.—PARTICLE MOTION AND RECORD CHARACTER FOR PLANE WAVES OF THE LONGITUDINAL AND TRANSVERSE TYPES

lengthwise (longitudinal) trace peak, whereas a trough and peak would coincide for the *SV* pulse.

As expected, field seismograms do not achieve the idealized simplicity of Figure 5. The approach is sufficiently close, however, that the identification principle is applicable.

In the preliminary stages of transverse-wave identification, other lines of evidence should converge with the particle-motion criteria to-

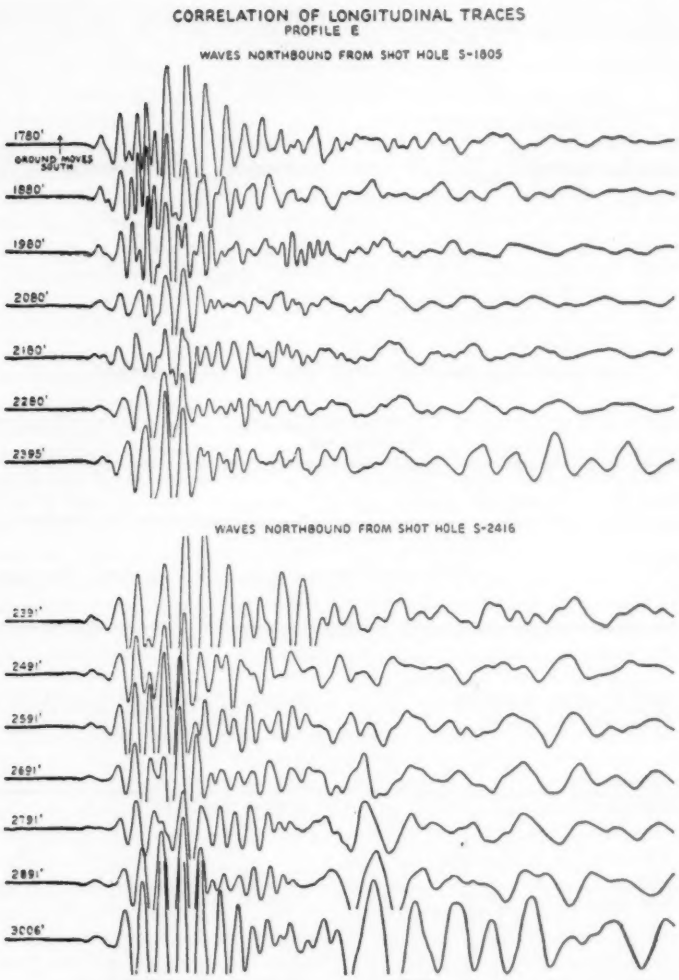


FIGURE 6.—CORRELATION OF LONGITUDINAL TRACES

ward a definitive result. For example: Poisson's ratio for most earth materials may be assumed to have a value between 0.0 and 0.4 which then requires that the ratio of longitudinal to transverse wave velocities lie between 1.41 and 2.45; a further hypothesis is that the thickness of seismic layers should be approximately the same when computed from the time-distance data of longitudinal and transverse waves. Thus, if the geology is known and the velocity layering is estimated, probable ray paths and travel-time curves can be constructed to test the most likely phase identifications.

Confirmation of phase identifications might also come from laboratory velocity measurements, but an objective of this work is to compare the field and laboratory velocity values, and these independent lines of information should not be entangled.

A final requirement in the identification of phases is that the record events, under consideration, be similar at different distances

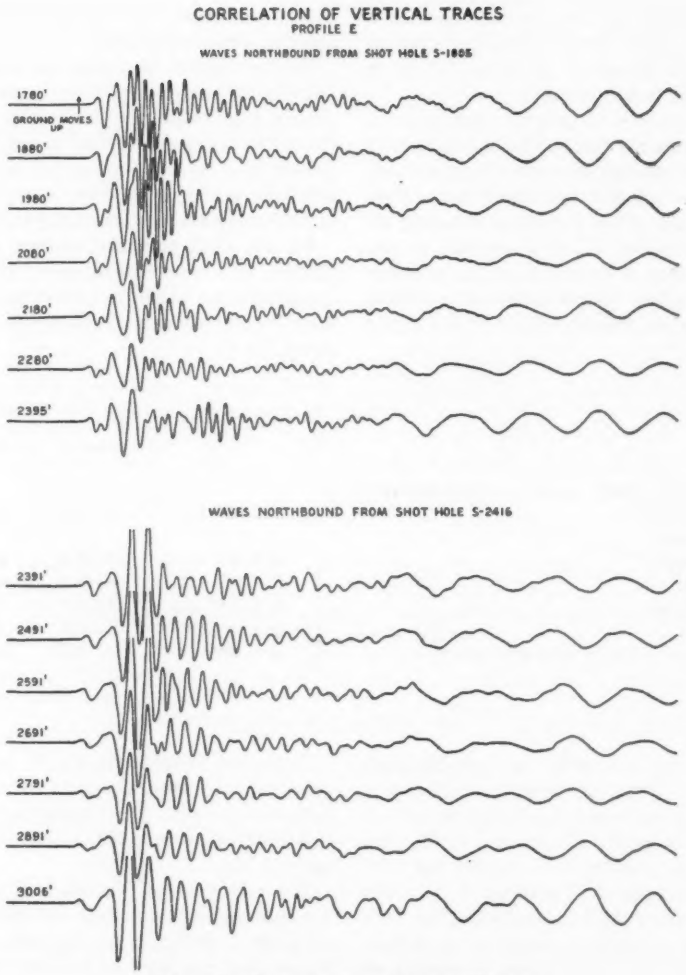


FIGURE 7.—CORRELATION OF VERTICAL TRACES

from the shot point. Figures 6 and 7 show the degree with which this requirement is met.

Correction to datum plane.—Irregularities in the observed travel times due to changes in the relative elevation of the shot and geophone and to changes in thickness or velocity of the surface-weathered zone were minimized by calculating a correction, T_w, to be subtracted from the observed travel time. The delay-time concept (Gardner, 1939; Nettleton, 1940) was used in all cases.

Observational Errors

Final velocity determinations were made by the method of least squares; these calculations also gave values of T_0, the intercept time at zero distance for each of the refracted waves.

If the data are subject only to random errors in the measurements of time and distance, the accuracy of the final result will depend on the number of time and distance observations for each phase. Thus, in this work, a sufficiently

large number of observations have been made so that the mean error of all observations for the principal phases P_1, S_1, P_2, and S_2 in no case exceeds ± 1.0 per cent.

This figure of ± 1.0 per cent would be useful for comparing the field and laboratory data, if it could be definitely shown that random errors, alone, have affected the observations. However, in view of the method of obtaining the surface correction, T_w, it is necessary to consider the effect of systematic errors; *maximum* values have been assigned to the errors arising from all known sources, and for phases P_2 and S_2 we find:

Source of error	Probable maximum error	Per cent error	
		Phase P_2	Phase S_2
Choice of pick point assuming proper correlation of record events	± 0.001 sec.	± 0.57	± 0.38
Reading record			
Shot instant	± 0.001 sec.	± 0.57	± 0.38
Pick point	± 0.001 sec.	± 0.57	± 0.38
Timing interval	± 0.0001 sec.	Trivial	Trivial
Distance	± 1′ in 3000′	Trivial	Trivial
Miscellaneous: variable lag in shot instant, etc.	± 0.0005	± 0.29	± 0.19
		± 2.00%	± 1.33%

To this must be added the probable maximum error of the weathering and surface correction, T_w. For phase P_2 the largest value of T_w does not exceed 5.7 per cent of the total travel time, and for phase S_2, 9.8 per cent. Then if the value of T_w is certain to be within ±20 per cent of the calculated value, the maximum error from this source in the travel time, corrected to the datum plane, is respectively ±1.14 per cent and ±1.96 per cent. Combining these errors with the previous estimates, it is found that the error in V_P can approach ±3.14 per cent and ±3.29 per cent for V_S. Errors in V_P and V_S will be slightly higher, but of the same order of magnitude.

In contrast, when the possibility of systematic errors is ruled out, the mean error of 18 observations of V_P is ±0.64 per cent and for 7 observations of V_S ±0.30 per cent.

Results

Phases P_1 and S_1.—Longitudinal velocities in the V_1 layer were obtained in directions parallel and perpendicular to the bedding with values agreeing within 0.5 per cent.

For the vertical subprofile, the least-square solution of 11 travel-time observations gave an average longitudinal vertical velocity of 9146 ft/sec. The observed vertical travel time for a distance of 211 feet from the top of the Neva limestone to the datum plane at an elevation of 1035 above sea level was 0.023^0 second.

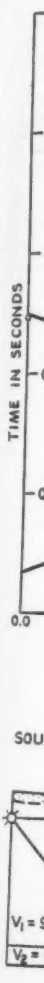
For the short horizontal profiles, a similar solution of 9 observations gave 9232 ft/sec. Assigning equal weight to both types of determination and averaging yields a value of 9189 ft/sec. for V_P.

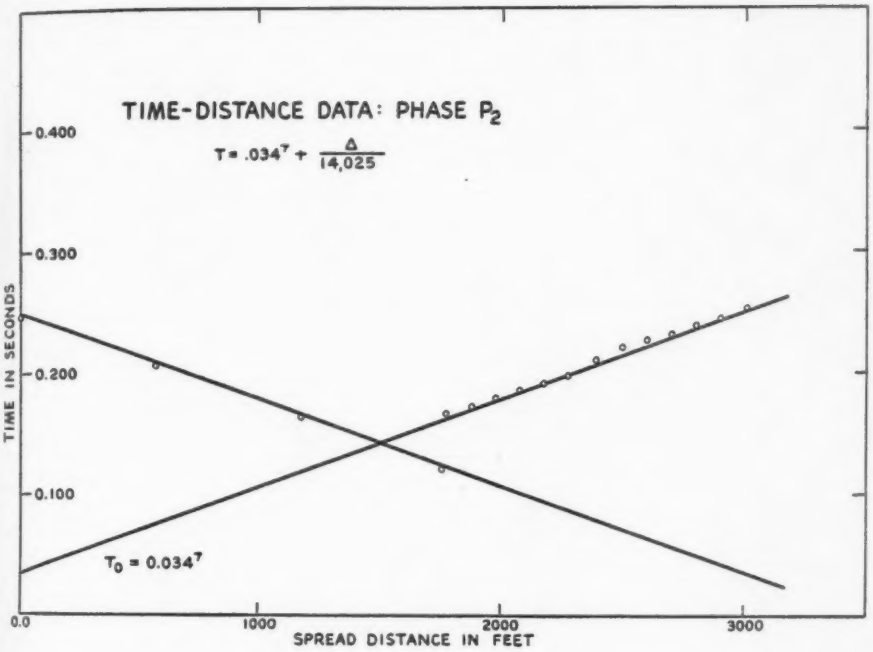
Transverse velocities in the V_1 layer were obtained only from the data of the short horizontal profiles. The least-square solution of 9 observations gave 5012 ft/sec. as the best value of V_S.

The record event identified as phase S_1 consisted of large-amplitude SV-type motion for which a distinctive turning point near the beginning of the phase could be correlated for all horizontal distances from 116 to 716 feet.

Phases P_2 and S_2.—The velocity of longitudinal waves in the Neva limestone, V_2 layer, was obtained from reversed horizontal profiles at spread distances up to 3032 feet, and, since the values of velocity and intercept time at zero distance are sensibly equal, the V_2 layer is shown to be horizontal.

For northbound waves, a least-square solution of 14 travel-time observations gave V_P as 13637 ft/sec. and $T_0 = 0.034^8$ second; for southbound waves, a least-square solution of 16 travel time observations at four different distances gave the corresponding values of





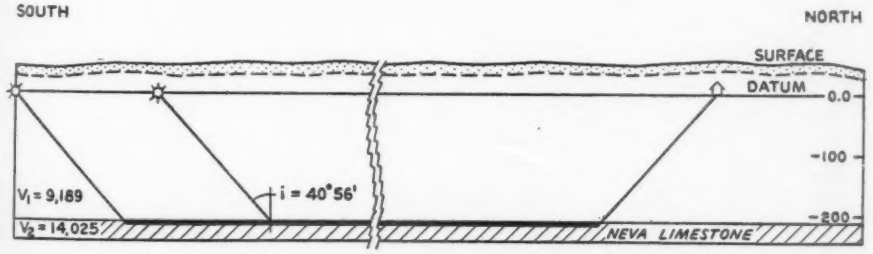


FIGURE 8.—TIME-DISTANCE DATA: PHASE P_2

14412 ft/sec. and 0.37^9 second; after averaging V_P = 14025 ft/sec. and $T_0 = 0.035^8$ second.

In the upper portion of Figure 8, the heavy lines indicate the travel time predicted for the wave path shown in the lower portion when V_P = 9189 ft/sec. and V_{P_2} = 14,025 ft/sec. For comparison, the observed travel times, reduced to the datum, are shown by the small circles. Coincidence of the observed and predicted travel times is the principal criterion for path identification.

A further verification can be obtained by converting the intercept time, T_0, to an equivalent vertical time and comparing with the directly observed vertical time from the datum to top of the Neva limestone. Thus, for a vertical distance of 211 feet, the equivalent vertical time as calculated from the average T_0 is 0.022^7 second, and the directly observed vertical time is 0.023^0 second.

Evidently, then, the top of the Neva limestone corresponds, at least for longitudinal waves, to the top of the seismic V_2 layer, but

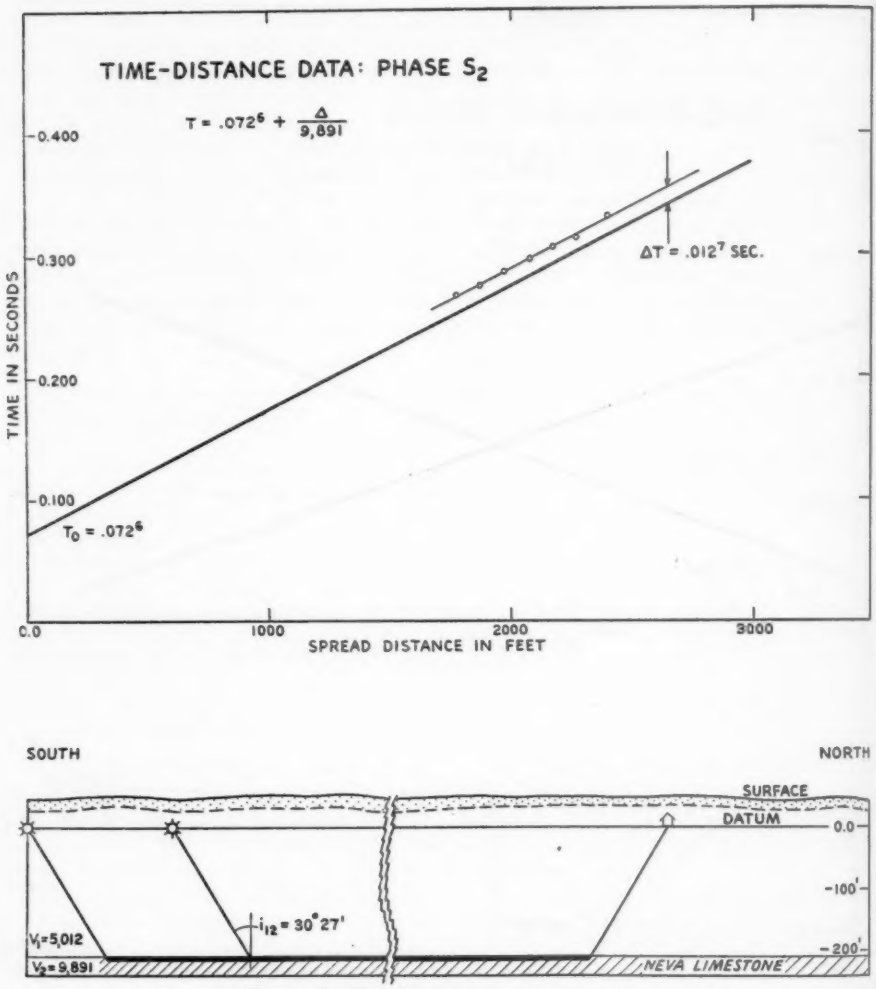


FIGURE 9.—TIME-DISTANCE DATA: PHASE S_2

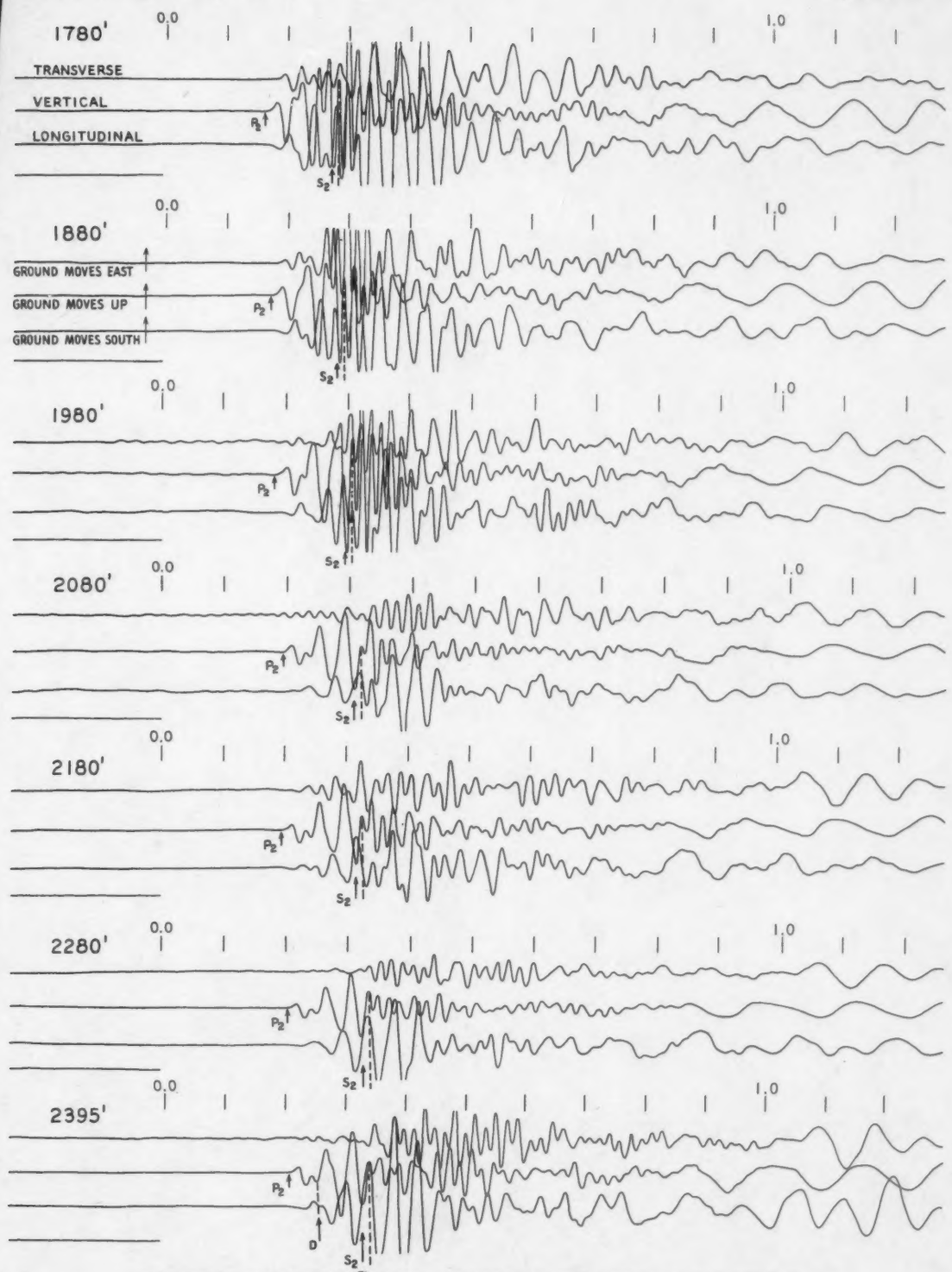
there need be no such correspondence for the base of the layers.

The event identified as phase P_2 can be seen in Plates 1–3.

Transverse waves refracted by the V_2 layer were studied on each of the longer horizontal profiles, but because of the large interval between the recording positions for southbound waves phase S_2 is difficult to identify; the closely spaced positions for northbound waves are better suited to this purpose, and for the latter a least-square solution of 7 travel-time observations gave $V_{S_2} = 9891$ ft/sec. and a corrected $T_0 = 0.075^3$ second.

In the upper portion of Figure 9, the heavy line indicates the travel times predicted for the wave path in the lower portion. For comparison, the observed travel times, reduced to the datum, are shown by small circles. Coincidence of observed and predicted times is the principal criterion for path identification.

The wave type was identified from a study



THREE-COMPONENT SEISMOGRAPH RECORDS: ENLARGEMENT OF PORTION OF PLATE 1

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of the recorded particle motion, as previously discussed. Since the true beginning of phase S_2 could not be constantly picked over large spread intervals, a later turning point, near the beginning of recorded transverse motion, was correlated over the interval 1780 to 2395 feet, and the times for this reference point, after reduction to the datum, were plotted as the small circles of Figure 9.

At distances of 1780, 1880 and 1980 feet, it is believed that the true beginning of the phase S_2 is recorded, and, in each case, pre cedes the reference turning point by 0.010 second. Thus the observed T_0, from the least-square solution of the seven travel time observations, is $0.085^3 - 0.010 = 0.075^3$ second as compared to the predicted T_0 of 0.072^6 second. This discrepancy is real and indicates the uncertainty in path identification; it can mean that the transverse wave was refracted at some horizon 8 feet lower than the top of the Neva limestone or that the true beginning of S_2 was not picked, or that the values of V_S and V_S are in error.

The event identified as phase S_2 can be seen in Plates 1 and 3. On each record the reference turning points used to obtain the velocity of S_2 are indicated by a broken vertical line; the beginning of phase S_2 is indicated by small arrows, preceding the reference mark by 0.010 second.

Other phases.—In addition to the phases of principal interest, several distinctive events which can be correlated through all or a large part of the spread interval from 1780 to 3006 feet are to be seen in Figures 6 and 7 and in Plates 1–3. These events may generally be identified as to wave type on the basis of their particle motion, and approximate velocities can be obtained by correlating distinctive reference points, but true beginnings are difficult to pick, hence the origin and wave paths are uncertain. Proper phase identification, specifying wave type and path, would require further field work.

In Plates 1–3, the event labelled "D" at distances of 2395 and 3006 feet can be correlated throughout the entire interval of 1780 to 3006 feet. It is approximately the beginning of the largest recorded motion and appears to be a steeply emergent longitudinal wave. From the study of time-distance graphs, it is found that event D has a travel time and velocity similar to that for phases P_1 and P_1P_1 at the larger distances of 3006, 2891, 2791, and 2691 feet, whereas at the shorter distances it arrives later than these phases and has a higher apparent velocity. If the surface corrections have not introduced a serious systematic error, it seems that event D does not represent the same phase at all spread distances. The largest recorded motion at these distances, however, is most probably due to a combination of the direct and reflected longitudinal waves.

Another event of interest is the steeply emergent transverse motion of SV type which arrives approximately 0.040 second after the reference point of phase S_2. At the larger spread distances this event has been labelled $S_3(?)$ in Plate 2. Within small limits of error its velocity is the same as that assigned to phase S_2, and, in fact, it may be a continuation of the S_2 motion rather than some later turning point associated with the transverse wave refracted at a deeper horizon.

From a travel time graph of the first motion recorded on the transverse trace, it is found that this event has a velocity essentially equal to P_2 but is later than P_2 in arrival time by 0.012 to 0.014 seconds. This delay is about 50 per cent smaller than would be expected for the transformed phase $S_1\ P_2\ P_1$; it is, however, almost exactly equal to the interval expected for a longitudinal wave transformed at the base of the weathered zone. The SH type motion may be attributed to irregularities of the base of this zone. Thus, it seems that significant quantities of energy were not supplied to SH type transverse waves at the shot, but that longitudinal waves incident on the irregular base of the weathered zone are partially transformed and lose a small part of their energy to SH type waves.

In the later parts of the records, long-period motions predominate, and arrival of the higher frequency "air wave" is in marked contrast to the otherwise regular motion. These long-period motions do not conform to the simple concept of a single Love and Rayleigh wave travelling at the surface.

Rayleigh wave motion with a velocity approximately equal to that of the air wave can

be seen at the end of each record. The particle appears to rotate in a retrograde sense, and the ratio of major to minor axes of the ellipse is highly variable.

Several reference points on the transverse trace can be correlated over appreciable distances to give velocities of 4500 and 3700 ft/sec. A reference point on the longitudinal trace gives a velocity of 3500 ft/sec. Apparent intercept times for these events range from 0.48 to 0.83 seconds, and, even with a liberal allowance for the uncertainty in picking the true beginning of these phases, it seems unlikely that they were generated at the shot point.

Of interest in connection with the amplitude of these long-period surface waves are the results of the vertical subprofile at S-0623 where the charge was detonated at or near the top of the V_2 layer. For seismometer locations near the hole, long-period motions, if they existed at all, were below the recording threshold; at greater distances these motions increased in amplitude but never became comparable to that found at the same distance, sensitivity, and charge size for shallow shot holes. In this respect surface waves from a small dynamite blast are qualitatively similar to those of a deep-focus earthquake.

Poisson's ratio.—By recognizing the possibility of a systematic error in the weathered zone correction, the field velocity values become:

$$V_{P_1} = 9189 \pm 300 \text{ ft/sec.}$$
$$V_{S_1} = 5012 \pm 170 \text{ ft/sec.}$$
$$V_{P_2} = 14{,}025 \pm 440 \text{ ft/sec.}$$
$$V_{S_2} = 9891 \pm 330 \text{ ft/sec.}$$

Poisson's ratio for the V_1 layer is thus:

$$0.24 < \sigma < 0.32$$

and for the V_2 layer[1]:

$$0.00 < \sigma < 0.11$$

By comparison with published values of Poisson's ratio (Birch *et al.*, 1942; Obert, Windes and Duvall, 1946; Windes, 1949) the value for the V_1 layer appears somewhat high, but not inordinately so when the extreme sensitivity of Poisson's ratio to slight changes in velocity is realized. For the V_2 layer, however, it is impossible to concede, on an *a priori* basis, that an unweathered, nonporous Permian limestone should have a true Poisson's ratio approaching zero.

It should be clearly understood that a significant contradiction exists, and that the field values of V_{P_2} and V_{S_2} would result in an erroneous identification of the material composing the V_2 layer.

To understand this contradiction, it is helpful to review the operations which produced it: the equipment and methods of field operation, the interpretation technique for wave paths, wave types, and velocities, and finally the theoretical relations used for calculating Poisson's ratio. Possibility of errors from the equipment or methods of field operation may be dismissed because of the close agreement for longitudinal waves between the actual and predicted depth of the V_2 refracting layer. To this extent, also, the interpretation techniques for the longitudinal wave path and velocity are eliminated as a source of appreciable error. The theoretical relation enabling Poisson's ratio to be calculated from longitudinal and transverse wave velocities assumes a homogeneous, isotropic material which probably has only an abstract existence, but this can hardly explain the contradiction since similar errors from this source have undoubtedly entered this and the previous published calculations. The remaining sources of error, the interpretation techniques for transverse wave paths or velocity cannot be so readily disposed of: particle motions characteristic of a steeply emergent SV type wave can be correlated at 100-foot intervals over a spread distance from 1780 to 2395 feet to yield a least-square solution of transverse wave velocity; the path of these waves is subject to greater uncertainty because the true beginning of this type motion must be distinguished from a background of large amplitude motion. If, as believed, the true beginning is unobscured at spread distances of 1780, 1880 and 1980 feet, the wave path is substantially correct, but the top of the refracting V_2 layer would be deeper by 8 feet for transverse waves than for longitudinal waves.

Most probably, then, the field values of V_{P_2} and V_{S_2} can be accepted as a reality, not the result of errors, but the very small value of Poisson's ratio is subject to rejection on the

[1] Excluding the possibility of a negative Poisson's ratio for these rocks.

basis that longitudinal and transverse velocities were not determined in the same material. The laboratory study of small core samples from the V_2 refracting horizon assists further resolution of this problem.

Laboratory Study

General Statement

The second phase of this work had as its objective the measurement of longitudinal and transverse wave velocities in small core samples of the Neva limestone, V_2 layer.

Since most of these observations were made at room temperature, and atmospheric pressure for specimens in the air-dry condition, it was necessary to investigate the effects of changing water content, pressure, and temperature in order that final values might be corrected to the probable conditions of the rock in place.

In an attempt to reduce the sampling error for this inhomogeneous limestone, 17 specimens from 8 horizons, within the bed, were studied. Of subsidiary interest are the results for 6 other specimens from 3 horizons in the V_1 layer.

Methods and Equipment

From a measurement of the resonant frequency of forced vibrations in a thin rod, values were obtained for both types of velocity. With the same equipment, the specimen could be excited in predominantly torsional modes from which the transverse velocity,

$$V_S = \sqrt{\mu/\rho}$$

was obtained, and then, by a simple change, in predominantly longitudinal modes from which the longitudinal velocity,

$$V_E = \sqrt{E/\rho}$$

was obtained. Where:

$V = 2fl$
V = the velocity of the wave
f = the resonant frequency of forced vibration
l = the length of the specimen
μ = the modulus of rigidity
E = Young's Modulus
ρ = the density

when it is assumed that the material under test is homogeneous and isotropic and when, for V_E, the specimen's length is considerably greater than its diameter.

The laboratory value of V_S in the thin rod is directly equivalent to that for a similar homogeneous, isotropic material of infinite extent. This is not true for longitudinal waves, however, and it is necessary to convert V_E to V_P by the relation:

$$V_P = V_E \sqrt{\frac{1-\sigma}{1-\sigma-2\sigma^2}}$$

where:

$V_P = \sqrt{\dfrac{k + 4/3\mu}{\rho}}$ = the longitudinal wave velocity in homogeneous, isotropic material of indefinite extent when k is the incompressibility modulus and μ is the rigidity modulus.

σ = Poisson's ratio = $\dfrac{\left(\dfrac{V_E}{V_S}\right)^2 - 2}{2}$

Subject to the above conditions, then, V_P and V_S may be calculated for the laboratory specimens. With additional data, these values may be empirically corrected to the probable conditions of water content, pressure, and temperature experienced by the rock, in place.

The equipment for this work was developed by Dr. Francis Birch in the study of metals, glasses, igneous and metamorphic rocks (Birch, 1937; Birch and Bancroft, 1938b; 1940; 1942). The equipment consists of a calibrated oscillator and power amplifier which supply a variable-frequency alternating current to a low-impedance, iron-core, driving coil. A small bar magnet, attached to one end of the test specimen by De Khotinsky cement, is positioned relative to the pole pieces of the driving coil so as to produce predominantly longitudinal or torsional vibrations, as desired; the strains in the test specimen probably do not exceed 10^{-6} and should therefore approximate the strains involved in the transmission of elastic waves, except in the vicinity of the charge. A similar magnet at the opposite end of the specimen generates voltage in an identical iron-core pickup coil, which is read, after amplification, with an oxide-rectifier type A.C. voltmeter. As the exciting frequency is varied, resonance is indicated by a sharp, nearly symmetrical deflection of the output meter.

The desirability of a supersonic transient method for velocity measurements in rocks was briefly investigated with the kind assistance of W. P. Mason at the Bell Telephone Laboratories (Mason and McSkimin, 1947). The results showed that vibrations with a wave length approaching the grain size or pore space of the rock are highly attenuated; samples of Neva limestone and Quincy granite showed a minimum loss of 16 db per centimeter for longitudinal carrier waves at 2.8 megacycles. Later studies on a finer-grained diabase showed considerably less loss. In view of these facts, the method seems promising for most rock types when the carrier frequency of the transient is 1.0 megacycles or less, and particularly when the specimen is under hydrostatic confining pressure. With appropriate crystals for transmitting and receiving, a direct measurement of V_P and V_S is possible.

Corrections

To reduce the disturbing effect on resonant frequency, caused by the small magnets cemented to the ends of the vibrating rod, small corrections were applied (Birch, 1937; Birch and Bancroft, 1938b).

In the case of longitudinal vibrations, a further correction was applied to the shorter test specimens to allow for the fact that the ratio of their diameters to lengths was not zero or some small number characteristic of a very long, thin rod (Bancroft, 1941).

Observational Errors

Summarizing for all air-dry test specimens, it was found that the resonant frequency of longitudinal vibration ranged from 4700 to 26,000 cps., and the total correction from 1.3 to 5.4 per cent, averaging about 2.5 per cent; the resonant frequency of torsional vibration ranged from 3000 to 19,400 cps., and the total correction from 0.3 to 1.8 per cent, averaging about 0.8 per cent. Thus, if the corrections, themselves, have an uncertainty of ±1 per cent, and the oscillator frequency calibration an uncertainty of ±0.5 per cent, the resonant frequency of longitudinal vibrations is known with an uncertainty of not more than 2.0 per cent, and for torsional vibrations not more than 1.0 per cent. The length of all specimens was measured to the nearest thousandth of an inch, and the error from this source may be neglected.

Thus it is estimated that, for any one determination of V_S and V_R, the velocity will be known within ±1 per cent and ±2 per cent, respectively. To obtain V_P, these uncertainties are multiply involved, first in the calculation of Poisson's ratio and second in the actual conversion of V_R to V_P. Thus Poisson's ratio, involving the square of both V_R and V_S, will be known to within ±6 per cent, and V_P will finally be known with an uncertainty of not more than ±4 per cent. It should be noted that these are estimates of the maximum uncertainty for one determination.

Results

Description of samples.—The core samples from shot hole S-0623 which have been studied in the laboratory may be grouped as samples of the V_1 or V_2 layer, and their significant properties described.

In the V_1 layer, three samples were studied; the results for the six test specimens cut from these samples are of minor interest because they do not approach a representative sample of the V_1 layer.

Sample 23: This 6-inch core of Crouse limestone was taken 4 feet from the base of the bed; it is a tan, crystalline, fossiliferous limestone with numerous solution cavities. Care was taken to avoid these cavities in cutting the test specimens.

Sample 5: This 7-inch core of the Cottonwood limestone was taken 4 feet from the base of the bed; it is a tan, coarsely crystalline limestone quite soft and easily cut.

Sample 7: This 9-inch core from the Eskeridge shale was taken 12.5 feet below the base of the Cottonwood limestone; the sample is a greenish-gray, fine-grained sand of poorly rounded, poorly sorted quartz grains in a clay matrix.

In the V_2 layer, 8 samples were studied; the results for the 17 test specimens cut from these samples are used as the basis for the comparison of field and laboratory velocity measurements. Core samples from the uppermost portion of the Neva limestone could not be obtained in lengths sufficient for laboratory measurement.

Sample 11: This 8-inch core was taken 15 feet from the base of the Neva limestone; it is an impure dark-gray, fine-grained, fossiliferous limestone. The interlocking, irregular carbonate grains, mostly less than 0.01 mm. in diameter, predominate in a matrix of clay; fossils are generally composed of coarser carbonate grains but may be associated with small amounts of chalcedony. No magnesium is detectable with conventional wet tests.

Sample 12: This 8-inch core was taken 14 feet from the base of the Neva limestone; it is a variant of sample 11, having fewer fossils and a more homogeneous appearance.

Sample 13: This 6-inch core was taken 13.5 feet from the base of the Neva limestone; it is a variant of sample 11, but darker because of a relatively greater abundance of clay matrix.

Sample 15: This 6-inch core was taken 11 feet from the base of the Neva limestone; it is a variant of sample 11, with but slightly greater abundance of clay matrix and fine carbonate grains which, on the whole, are slightly coarser.

Sample 16: This 9-inch core was taken 9 feet from the base of the Neva limestone; it closely resembles sample 13, but the clay matrix is in relatively greater abundance and the fine carbonate grains are nearly all surrounded by a band of clay.

Sample 18: This 7-inch core was taken 6.5 feet from the base of the Neva limestone; it is strikingly different from all other samples in that it is largely composed of angular carbonate pebbles up to 1 inch in the longest dimension. The fine-grained pebbles are well cemented by a coarse-grained carbonate binder to produce a hard dense rock. Clay or silty material is absent, and thus the carbonate grains are in interlocking mutual contact. This is the only specimen in which magnesium was definitely detected by means of a wet test.

Sample 19: This 7-inch core was taken 2.5 feet from the base of the Neva limestone; it is similar to sample 16.

Sample 20: This 6.5-inch core was taken 0.5 feet from the base of the Neva limestone; it is a variant of sample 16.

From each sample, at least two test specimens were cut and carefully machined to the form of a right circular cylinder. The dimensions of the final core pieces can be seen in Table 1 which also includes the air-dry densities, calculated from the dimensions and mass of each piece.

The coring process gave samples with their largest dimension perpendicular to the bedding. Laboratory velocities are thus measured in a direction perpendicular to the bedding, while the direction of field measurements paralleled the bedding.

Summary of air-dry velocities.—Since the water content of a core sample may change appreciably, by absorption of the drilling fluid or by subsequent evaporation to the air, the standard reference for all samples was arbitrarily taken as its air-dry condition at room temperature and atmospheric pressure. Thus laboratory velocities were measured under these conditions, and empirical corrections used to reduce these values to the most probable ones for the rock, in place.

TABLE 1.—DIMENSIONS AND AIR-DRY DENSITIES OF SPECIMENS

Core piece	Radius (cm.)	Length × 2 (cm.)	Air dry density (gm./cc.)
23A	.796	31.61	2.39
23B	.774	26.51	2.35
5A	.790	44.07	2.54
5B	.787	45.57	2.52
7A	.762	19.16	2.15
7B	.795	28.05	2.19
11A	.788	25.44	2.64
11B	.793	19.57	2.67
11C	.686	27.66	2.66
12A	.771	34.30	2.64
12B	.768	33.63	2.64
13A	.783	16.12	2.53
13B	.785	16.59	2.51
15A	.796	27.40	2.60
15B	.794	22.98	2.60
16A	.766	42.12	2.51
16B	.783	26.95	2.52
18A	.789	33.58	2.66
18B	.792	19.95	2.67
19A	.794	11.18	2.49
19B	.794	12.45	2.50
20A	.795	19.01	2.66
20B	.771	22.00	2.64

The results for the three samples from the V_1 layer are summarized in Table 2. Values for the core pieces cut from the same sample generally agree within the estimated maximum uncertainty of ±2 per cent for V_R, ±1 per cent for V_S, and ±6 per cent for Poisson's ratio. The velocities of the Crouse limestone (sample 23) are similar to those for the Neva limestone V_2 layer; this sample, though near the surface, has not been weathered in the

seismic sense. Velocities for the Cottonwood limestone (sample 5) are low because of its coarsely crystalline structure, and comparable to that for the greenish-gray sand of sample 7; Poisson's ratio for these two rock types is quite different, however, and reflects, among other things, the extreme sensitivity of this quantity to relative changes in V_S and V_P.

TABLE 2.—SUMMARY OF RESULTS FOR V_1 LAYER
Specimens: Air-Dry; Room Temperature; Atmospheric Pressure

Core piece	V_R (ft/sec)	V_S (ft/sec)	V_P (ft/sec)	σ
23A	13,806	8,789	14,908	0.234
23B	13,438	8,569	14,465	.230
5A	7,493	4,688	8,438	.277
5B	7,205	4,541	7,969	.260
7A	7,841	5,302	7,913	.094
7B	7,933	5,374	8,005	.089

The results for the eight samples from the Neva limestone, V_2 layer, are summarized in Table 3. The values for core pieces cut from the same sample generally agree within the estimated maximum uncertainties; there are, however, a few exceptions, most probably explained on the basis of slightly different contents of water in the air-dry core pieces. It is entirely possible, however, that one piece has been fractured or otherwise more severely maltreated in the original coring or later machining than has its companion; significant changes in composition or aggregation character, within a single sample, are less likely.

Perhaps the most striking feature of the data is their range of variation. Thus, for closely spaced samples, all within the Neva limestone, V_S ranges from 10,600 to 8100 ft/sec., V_P from 21,000 to 12,800 ft/sec., and Poisson's ratio from 0.328 to 0.116. Such changes are in excess of the accidental differences, and far greater than the experimental error; they must then reflect real differences in the material.

It is not our objective to study the nature of these differences in any detail, but certain tentative conclusions may be of interest. First, with regard to composition, samples 16 and 13 with low values of Poisson's ratio and low velocities have the greatest relative amounts of clay matrix, while sample 18 with the least clay matrix and the only detectable amount of magnesium has consistently high velocities and the highest value of Poisson's ratio. Thus the minor constituents of the limestone significantly influence its elastic properties. In this case, however, the minor constituents may not, alone, contribute to the differences in elasticity because the density of these samples correlates in precisely the same manner, and porosity may therefore be a contributing factor.

TABLE 3.—SUMMARY OF RESULTS FOR V_2 LAYER
Specimens: Air-Dry; Room Temperature; Atmospheric Pressure

Core piece	V_R (ft/sec)	V_S (ft/sec)	V_P (ft/sec)	σ
11A	16,329	10,272	18,140	0.264
11B	17,227	10,551	21,138	.334
11C	16,676	10,335	19,419	.303
12A	16,552	10,256	19,275	.303
12B	16,571	10,259	19,327	.304
13A	12,602	8,435	12,792	.116
13B	12,191	8,159	13,375	.116
15A	15,610	9,951	16,814	.231
15B	15,853	9,928	17,818	.275
16A	13,231	8,789	13,504	.133
16B	13,753	8,940	14,331	.184
18A	17,369	10,663	21,004	.327
18B	17,116	10,502	20,741	.328
19A		8,924		
19B		8,067		
20A	14,531	9,249	15,689	.234
20B	13,553	8,825	14,094	.180

Thus, if the samples are pictured as a collection of very fine calcium carbonate grains in a matrix of even finer clay minerals, it is reasonable to expect the aggregate to offer the most resistance to compressional or shearing stresses when the carbonate grains are in interlocking, mutual contact and therefore exhibit high velocities when porosity is small and the clay matrix absent.

Returning to the values of Table 3, it is appropriate to inquire how the most representative velocities may be obtained from the variable data. Since these variations are real and not attributable to accidental errors, we are faced with the same problem which confronted the seismic waves in the field, and it would be desirable to "average" the variations in exactly

the same way. As previously mentioned, however, the available theoretical solutions of this problem are inadequate, and we must rely on general measures of central tendency.

An instructive step is the plotting of frequency curves or histograms from which the dispersion of the data and its central tendency can be judged. Thus by choosing cell sizes of 500 and 2000 ft/sec., in proportion to the relative uncertainties in V_S and V_P, the following velocity distributions are obtained.

For V_S:

Velocity in Ft./Sec.	Freq.	Identity of Specimens	Air-Dry Density (Gms./CC)
8,000– 8,500	3	13A, 13B, 19B	2.53, 2.51, 2.50
8,500– 9,000	4	16A, 16B, 19A, 20B	2.51, 2.52, 2.49, 2.64
9,000– 9,500	1	20A	2.66
9,500–10,000	2	15A, 15B	2.60, 2.60
10,000–10,500	4	11A, 11C, 12A, 12B	2.64, 2.66, 2.64, 2.64
10,500–11,000	3	11B, 18A, 18B	2.67, 2.66, 2.67

For V_P:

Velocity in Ft./Sec.	Freq.	Identity of Specimens	Air-Dry Density (Gms./CC)
11,000–13,000	1	13A	2.53
13,000–15,000	4	13B, 16A, 16B, 20B	2.51, 2.51, 2.52, 2.64
15,000–17,000	2	15A, 20A	2.60, 2.66
17,000–19,000	2	11A, 15B	2.64, 2.60
19,000–21,000	4	11C, 12A, 12B, 18B	2.66, 2.64, 2.64, 2.67
21,000–23,000	2	11B, 18A	2.67, 2.66

This tabulation shows bimodal distributions for V_S and V_P which are strikingly similar. It should be noted that the maxima of each distribution are composed of values for the same core pieces, and that the maxima are separated by significant amounts: approximately seven times the maximum experimental error in V_S and four times the maximum experimental error in V_P. Evidently we are dealing with two materials of essentially different properties, and thus on the basis of density and laboratory velocity determinations the Neva limestone may be subdivided into two distinct types of alternating microlayers with thickness of the order of 1 foot.

The first of these materials, designated as Neva A, has modal velocity values of:

$$8500 < V_S < 9000 \text{ ft/sec.}$$
$$13{,}000 < V_P < 15{,}000 \text{ ft/sec.}$$

and the second, designated as Neva B:

$$10{,}000 < V_S < 10{,}500 \text{ ft/sec.}$$
$$19{,}000 < V_P < 21{,}000 \text{ ft/sec.}$$

The material of the Neva A microlayers is characterized by lower elasticities, lower densities, darker color due to greater abundance of clay matrix, somewhat coarser irregular carbonate grains, mostly less than 0.01 mm. in diameter, with fewer interlocking mutual contacts and a greater abundance of carbonate grains entirely surrounded by a band of extremely fine-grained silt. The material of the Neva B microlayers has higher elasticities, higher densities, lighter color, less clay matrix, and finer irregular carbonate grains in interlocking mutual contact. Neva B material may also be higher in magnesium content, but simple qualitative tests were not of sufficient sensitivity to assure this conclusion.

These modal velocities, then, may be taken as the central tendency values for the laboratory measurements on air-dry specimens.

Variables affecting velocity.—In order to reduce the air-dry velocity measurements to the most probable values for the rock, in place, the influence of changing water content, pressure, and temperature has been briefly studied for selected specimens principally from the Neva limestone, V_2 layer. It will be shown that the air-dry longitudinal velocities should be reduced about 4 per cent and the air-dry transverse velocities about 6 per cent to correspond with the saturated bedrock environment. Corrections for the pressure are probably less than 1 per cent, and corrections for temperature are negligible.

Effect of water content on velocity: Summarized in Table 4 are the effects of increasing and decreasing water content on the various velocities and Poisson's ratio of eight test specimens from six samples; except for

the greenish-gray sand (core pieces 7A and 7B) all are from the Neva limestone, V_2 layer.

The data are classed as Dried specimens, Air-dry specimens, and Saturated specimens or air-dry specimen.[2] Since V_S and V_R do not decrease at the same rate, Poisson's ratio changes, and for most specimens it is increased by the addition of water.

TABLE 4.—EFFECT OF WATER CONTENT ON VELOCITIES AND POISSON'S RATIO
Specimens: Room Temperature and Atmospheric Pressure

Core piece	Dried specimens					Air-dry specimens				Saturated specimens				
	$-P$	V_R	V_S	V_P	σ	V_R	V_S	V_P	σ	$+P$	V_R	V_S	V_P	σ
	Per cent	(ft/sec)	(ft/sec)	(ft/sec)		(ft/sec)	(ft/sec)	(ft/sec)		Per cent	(ft/sec)	(ft/sec)	(ft/sec)	
7A	0.58	8,225	5,607	8,277	0.076	7,841	5,302	7,913	0.094					
7B	0.44	8,163	5,761	8,163	0.0027	7,933	5,374	8,005	0.089					
11A	0.09	16,584	10,377	18,674	0.277	16,329	10,272	18,140	0.264					
11C	0.16	16,788	10,410	19,475	0.300	16,676	10,335	19,419	0.303	0.48	14,901	9,278	17,060	0.290
12A						16,552	10,256	19,275	0.303	0.57	15,863	9,787	18,783	0.314
15A						15,610	9,951	16,814	0.231	0.82	14,983	9,564	16,096	0.228
16B						13,753	8,940	14,331	0.184	1.73	12,664	8,120	13,487	0.217
18B						17,116	10,502	20,741	0.328	0.27	16,909	10,285	21,456	0.351

$-$ P = Water removed from air-dry specimen in per cent by weight.
$+$ P = Water added to air-dry specimen in per cent by weight.

according to the laboratory treatment prior to measurement. To obtain the conditions for dried specimens, each air-dry piece was placed in a sealed tube and evacuated with an oil-type pump for successive periods of 18 hours at a temperature of 70°C until reaching a constant weight; the pumping time per sample did not exceed 54 hours. The air-dry condition is the original state of the material on arrival at the laboratory. The saturated condition was achieved by fully immersing each piece in a water-filled container and evacuating over the water with an aspirator for 59 hours. In this latter case, it is quite important that the saturated specimen come to equilibrium with the added water before attempting velocity determinations; each was permitted to stand for a minimum of 24 hours in a sealed tube, at room temperature, before measurements were made.

For core piece 11C the sum of the water added (+P) and water removed (−P) is 0.64 per cent, an approximate measure of porosity on the basis of weight, or on a volume basis, about 1.7 per cent.

Velocities in Table 4 are decreased for most samples in an amount ranging from 2 to 12 per cent, by the addition of water to the dried

Core piece 11C was selected for further study, and in this instance the air-dry specimen was dried and then fully saturated in accordance with the above procedures; following this the water was removed in gradual steps by short intervals of low-pressure evacuation, and after each equalization period, velocities measured. The results may be seen in Figure 10, where the ordinate is per cent change in velocity and the abscissa is per cent, by weight, of water added or removed from the air-dry specimen. The heavier line applies to V_S, and the lighter to V_R. It should be noted principally that the addition of small quantities of water to the dry or air-dry specimen may reduce the velocities considerably.

As water is gradually removed, velocities do not increase appreciably until the specimen is almost in its original dried condition. From the shape of this curve, it seems likely that complete data for V_R and V_S as a function of water content would describe a "hysteresis curve" with nearly constant maximum and

[2] Velocities of all core pieces except 18B decrease with increasing water content, and Poisson's ratio increases for all except 11A, 11C, and 18B. These exceptions are attributable to the experimental error which is of the same magnitude as the change for these specimens.

minimum values of velocity. Actually the velocities at the extremes of saturation and dryness may not be precisely constant because the density of the material is changing slightly and the velocities are determined by the relations $V_S = \sqrt{\mu/\rho}$ and $V_R = \sqrt{E/\rho}$; the

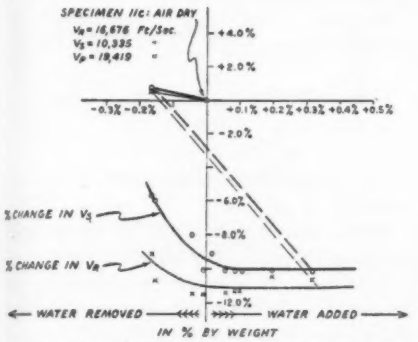

FIGURE 10.—VARIATION OF VELOCITIES WITH WATER CONTENT

After drying and abruptly saturating specimen 11C, velocities were determined in successive steps as the water was removed. The heavier line records the per cent change in V_S as a function of water added or water removed from the air-dry specimen; the lighter line applies to V_R.

change due to this effect, however, would be minor, indeed, by comparison with the variation due to change in the rigidity modulus or Young's modulus of the specimen (Born and Owen, 1934–1935).

Thus for the low-porosity Neva limestone samples an average addition to the air-dry specimen of 0.8 per cent, by weight, of water is sufficient to achieve saturation, and, on the average, this reduced V_S by 6.4 per cent and V_P by 4.3 per cent. Laboratory velocities should seemingly then be reduced by these amounts to correspond with this element of the bedrock environment since the Neva limestone in this area is a minimum of 100 feet below the level of the permanent water table.

EFFECT OF HYDROSTATIC CONFINING PRESSURE ON VELOCITY: If the average density of the V_1 layer is taken as 2.3 gms./cc. the hydrostatic pressure at the top of the Neva limestone, V_2 layer, may be estimated as 250 lbs./sq. inch. We may expect this small amount of pressure to have very little effect on the velocities in a dense, low-porosity material, and such a prediction was verified for V_S; certain experimental difficulties (Birch, 1937), however, obstruct the measurement of V_R under pressure.

Sample 18A was jacketed with thin copper foil to exclude the pressure medium from the pores of the rock, and V_S obtained at various values of hydrostatic confining pressure. The results appear in Table 5.

TABLE 5.—EFFECT OF HYDROSTATIC CONFINING PRESSURES ON THE TRANSVERSE VELOCITY OF SPECIMEN 18A

Pressure in lbs./sq. in.	Per cent increase in V_S relative to value at atmospheric pressure
800	1.60
1,500	1.63
1,800	1.69
4,000	2.09

All measurements were made at room temperature. The pressure readings are uncertain by ± 10 per cent.

Thus, by interpolating to a pressure of 250 lbs./sq. in. it is concluded that V_S was increased about 0.5 per cent; the increase in V_P will almost certainly be of the same order of magnitude. But since specimen 18A is more dense than the average sample of Neva limestone, it is possible that samples such as 16 might have their velocities increased by as much as 1 per cent. This figure, then, can be applied as a further correction for the air-dry velocity values measured at atmospheric pressure, although no conclusion would be altered if the pressure corrections were neglected.

EFFECT OF TEMPERATURE ON VELOCITY: The temperature at a depth of 248 feet will have a most probable value not far different from the mean annual surface temperature which in turn is about the same as the room temperature, 21°C, at which the air-dry measurements were made. Evidently the temperature correction will be quite small, since elastic wave velocities in rock are relatively insensitive to temperature changes (Ide, 1937; Birch, 1943).

A study of core piece 11A shows that the decrease in V_S and V_P for a temperature increase of 30C° was much less than the maximum experimental error; the computed value of change in V_S per centigrade degree of temperature change was about 2 ft./sec. and for

V_P about 4 ft./sec.; they may certainly be neglected. In contrast, a brief study of sample 7, the greenish-gray sand from the V_1 layer, shows relatively greater sensitivity to temperature change and may approach -1 per cent per 100C° for V_S and slightly more for V_P.

EFFECT OF MEASUREMENT FREQUENCY AND AMPLITUDE ON VELOCITY: If the rocks of the V_2 layer were highly dispersive, a change in the amplitude or frequency of elastic body waves in this material would be accompanied by a change in propagation velocity. Then since the field measurements were made at frequencies in the band from 20 to 70 cps. and the laboratory measurements at several thousands of cycles per second, the two results might not be comparable even though the amplitude of motion was similar.

A laboratory measurement of velocities over a range from 20 to 40,000 cps. is impractical, but by comparing the fundamental resonant frequency of a test specimen with its harmonic frequencies and by observing that resonant frequency is independent of driving voltage over a 10 to 1 range, it is concluded that samples of the V_2 layer are not dispersive in the frequency and amplitude range of laboratory measurements. A considerable band between 3000 and 20 cps. remains to be investigated.

Dissipation constant Q for materials of the V_2 layer.—The internal friction of a material disturbed by elastic waves acts to reduce or damp the amplitude of the disturbance. This action, in combination with other factors, determines the amplitude of a seismic wave at any particular distance from its source.

The dissipation "constant", Q may be taken as a measure of the internal friction or damping. Metals and glasses generally have a high Q and small internal friction; rocks have a lower Q and appreciably greater damping; extremely low values of Q are probable in unconsolidated materials and in the seismically weathered zone. The quantity Q is determined from the shape of the resonance curve of amplitude versus the frequency for forced vibrations in the thin rods by the relation:

$$Q = \frac{f_o}{\Delta f}$$

where: f_0 = the resonant frequency of forced vibrations, and Δf = the band width of the resonance curve at amplitudes equal to 0.707 times that at f_0.

Thus for the air-dry specimens with the most nearly symmetrical resonance curves the following values of Q were obtained at room temperature and atmospheric pressure:

Core piece	Q for torsional mode	Q for longitudinal mode
15A	440	—
18A	690	510
16A	230	—

The above values are uncertain by approximately ± 10 per cent.

With a change from the air-dry to the dried condition the value of Q increases slightly; on saturation, Q decreases very considerably, but the shape of the resonance curve alters so that the amount is uncertain.

Final values.—On the basis of these empirical studies, water content and hydrostatic confining pressure are the only independent variables exerting a significant influence; they were shown to produce opposite effects of relatively small magnitude on both V_S and V_P. Thus in correcting the laboratory measurements for specimens in the air-dry condition under atmospheric pressure to the most probable values for the same materials when saturated with water and under a hydrostatic confining pressure of 250 lbs./sq. inch, a total correction of -5 per cent should be applied for V_S, and -3 per cent for V_P.

COMPARISON AND CONCLUSION

At this stage the modal velocity values from laboratory determinations on Neva A material:

$$8100 < V_S < 8600$$
$$12{,}600 < V_P < 14{,}500 \text{ ft/sec}$$

and Neva B material:

$$9500 < V_S < 10{,}000$$
$$18{,}400 < V_P < 20{,}400 \text{ ft/sec}$$

may be compared with the limiting values found for the field measurements:

$$9600 < V_S < 10{,}200 \quad 13{,}600 < V_P < 14{,}500 \text{ ft/sec}$$

It is seen, at once, that the wide range of laboratory values encompasses the limiting

values of the field measurements, and to this restricted degree there is agreement between field and laboratory determinations. Order of magnitude differences do not exist between the two types of studies.

In detail, however, the assigned limits of error do not permit the field values of V_S and V_P to be matched with corresponding values for Neva A or Neva B. Or in terms of Poisson's ratio

$$0.06 < \text{Laboratory determination of Poisson's ratio for Neva A material} < 0.27$$

$$0.29 < \text{Laboratory determination of Poisson's ratio for Neva B material} < 0.36$$

$$0.00 < \text{Field determination of Poisson's ratio}^3 < 0.11$$

Thus the field determinations cannot be said to show detailed agreement with laboratory values for either type microlayer material.

An examination of this disagreement shows that the laboratory determinations of Poisson's ratio for the Neva A and B materials are comparable with previous studies on limestone (Birch *et al.*, 1942; Obert, Windes, and Duvall, 1946; Windes, 1949) while the field value is in marked disagreement. As previously suggested, at least two possibilities could produce acceptable measurements of field velocities, but an unacceptable value of Poisson's ratio:

(1) Since the Neva limestone refracting horizon contains thin, alternating layers of different density and elasticity, the recordable energy of field-generated longitudinal and transverse body waves may have travelled along this horizon at "averaged" velocities different from the modal values for the materials of the component microlayers, or

(2) In the same circumstance, the recordable energy of longitudinal body waves may have travelled with a velocity determined by the density and elastic parameters of the Neva A microlayer materials while the Neva B microlayer materials determined the velocity of transverse body waves.

The data of this work substantiate the latter hypothesis since for laboratory measurements:

$$9500 < V_S \text{ (Neva B)} < 10{,}000$$
$$12{,}600 < V_P \text{ (Neva A)} < 14{,}500 \text{ ft/sec.}$$

and for field measurements:

$$9600 < V_S < 10{,}200 \quad 13{,}600 < V_P < 14{,}500 \text{ ft/sec.}$$

Or in terms of Poisson's ratio the cross-correlated laboratory measurements yield[4]:

$$0.00 < \text{Poisson's ratio} < 0.12$$

as compared to the field measurement:

$$0.00 < \text{Poisson's ratio} < 0.11$$

As a consequence, it appears that certain sedimentary rocks may be considered homogeneous and isotropic only after subdivision into small rock units whose thickness will be small in comparison with most seismic wave lengths as well as the customary geological subdivisions of a rock column.

Inasmuch as this study casts no doubt on the basic assumption that field and laboratory determinations of elastic parameters will be in good agreement for homogeneous, isotropic materials under the same environmental conditions, the identification of buried sedimentary rock seems possible with seismic field instruments and techniques specifically designed to deal with waves refracted by thin layers.

In summary, it is concluded that:

(1) Transverse elastic body waves of the *SV* type can be generated by a small dynamite blast in sedimentary rocks if the shot hole is shallow and high confining pressures are avoided.

(2) The velocity of an *SV* type body wave can be determined with approximately the same accuracy as a longitudinal type body wave. The transmission path is subject to greater uncertainties.

(3) Among the independent variables affecting velocities in a near-surface, Permian limestone, water content is the most important. Pressure, temperature, and measurement direction, amplitude, or frequency have relatively minor effects.

(4) The Neva limestone may be subdivided on the basis of density and laboratory determinations of velocity into two distinct types of alternating microlayers with thickness of the order of 1 foot.

(5) Under the conditions of this work, elastic parameters and Poisson's ratio computed from

[3] Excluding the possibility of a negative Poisson's ratio for these rocks.

[4] Excluding the possibility of a negative Poisson's ratio for these rocks.

field measurements differ significantly from values determined by laboratory measurement.

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1905 National Bank Building, Tulsa, Oklahoma

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